# Bulletin
# of
# The Geological Society of America

VOLUME 61 November 1950 NUMBER 11

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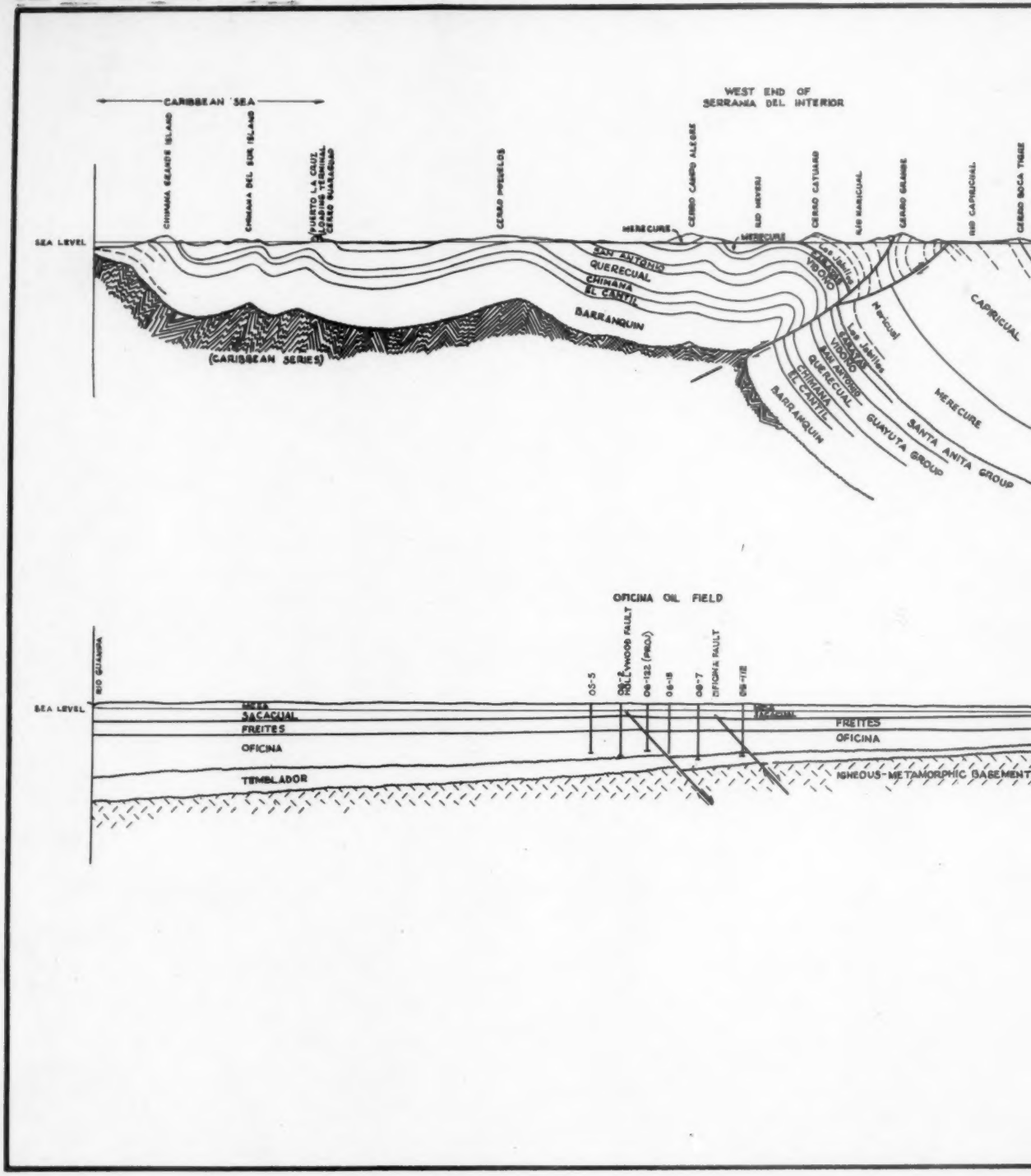
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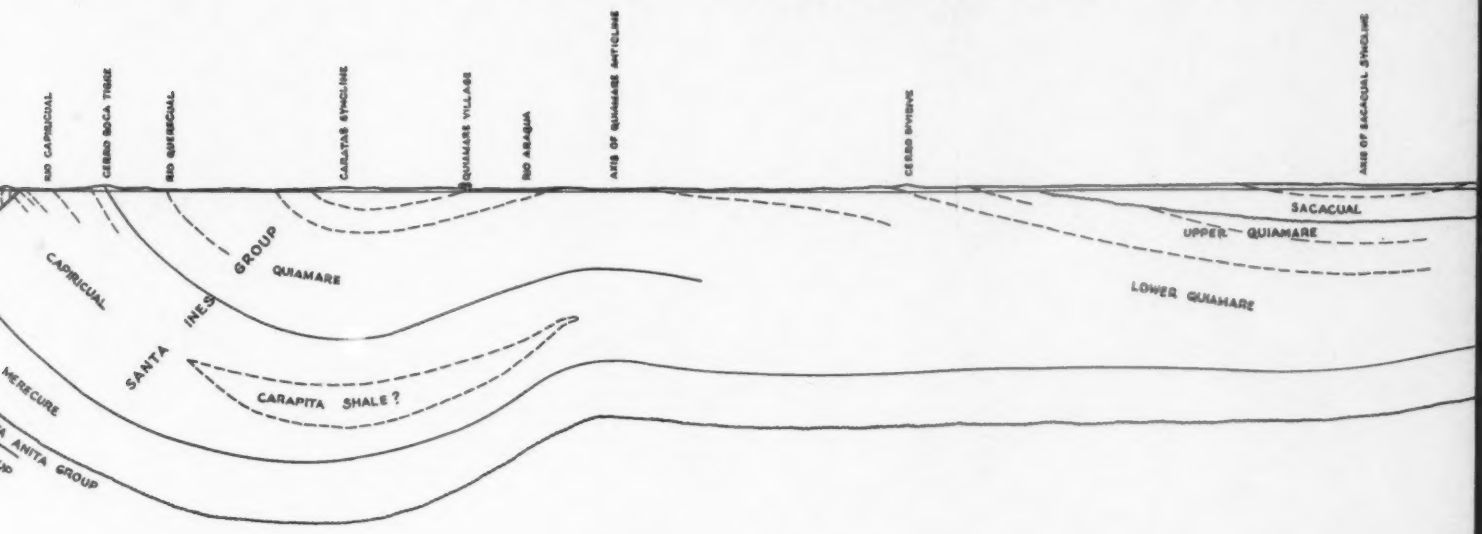


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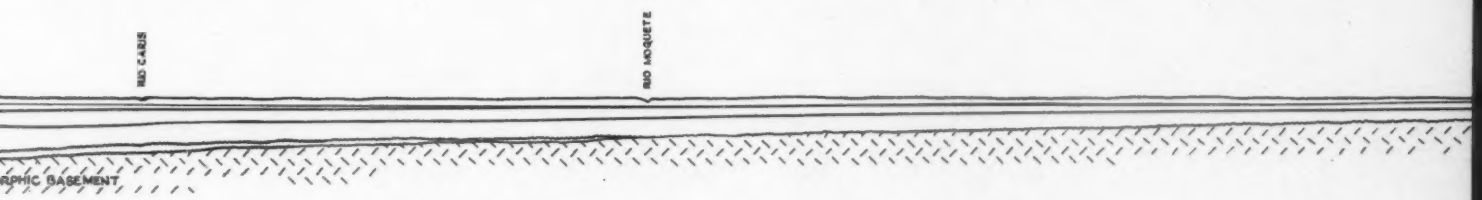


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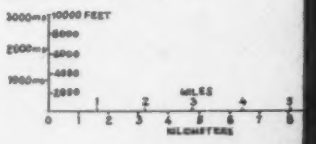


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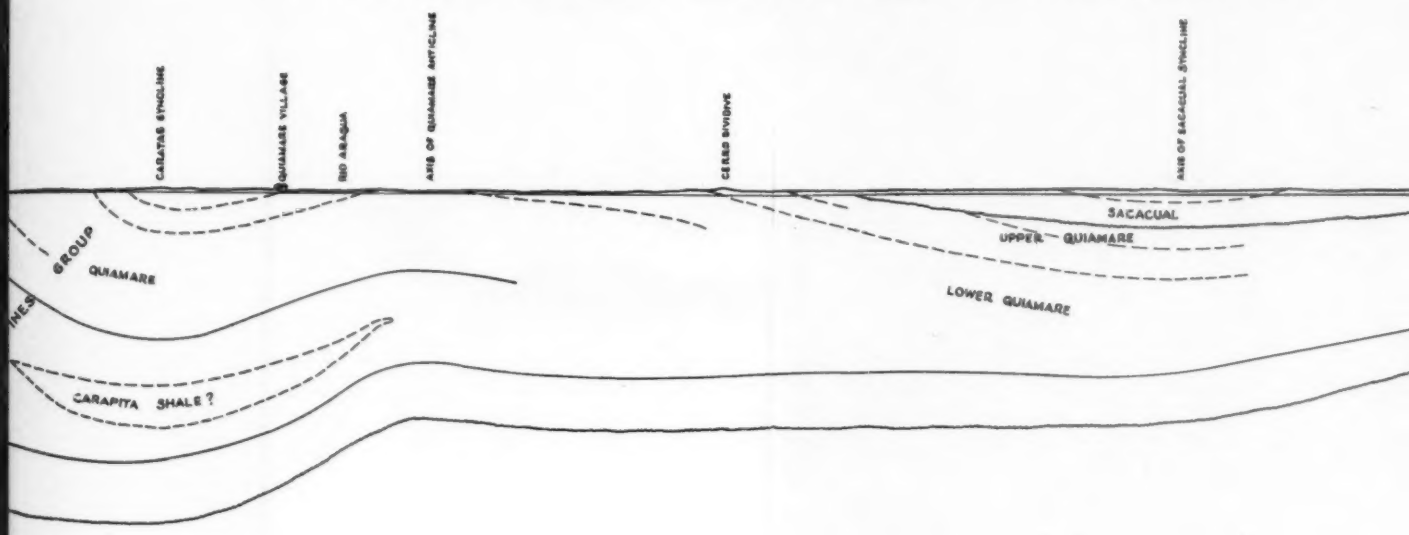


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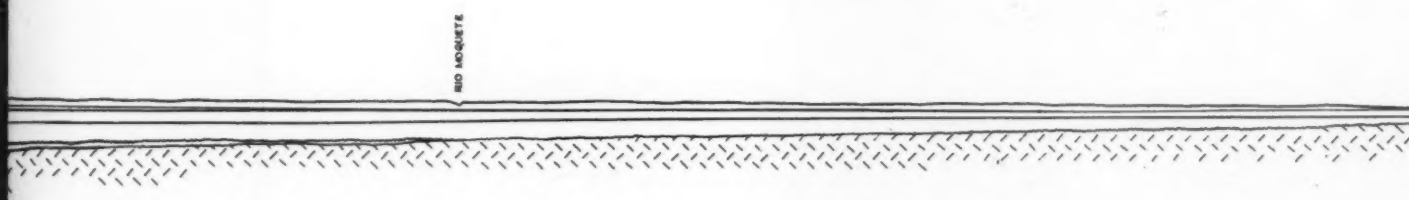


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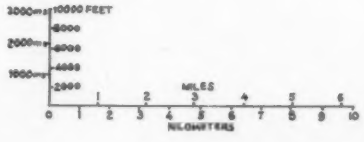


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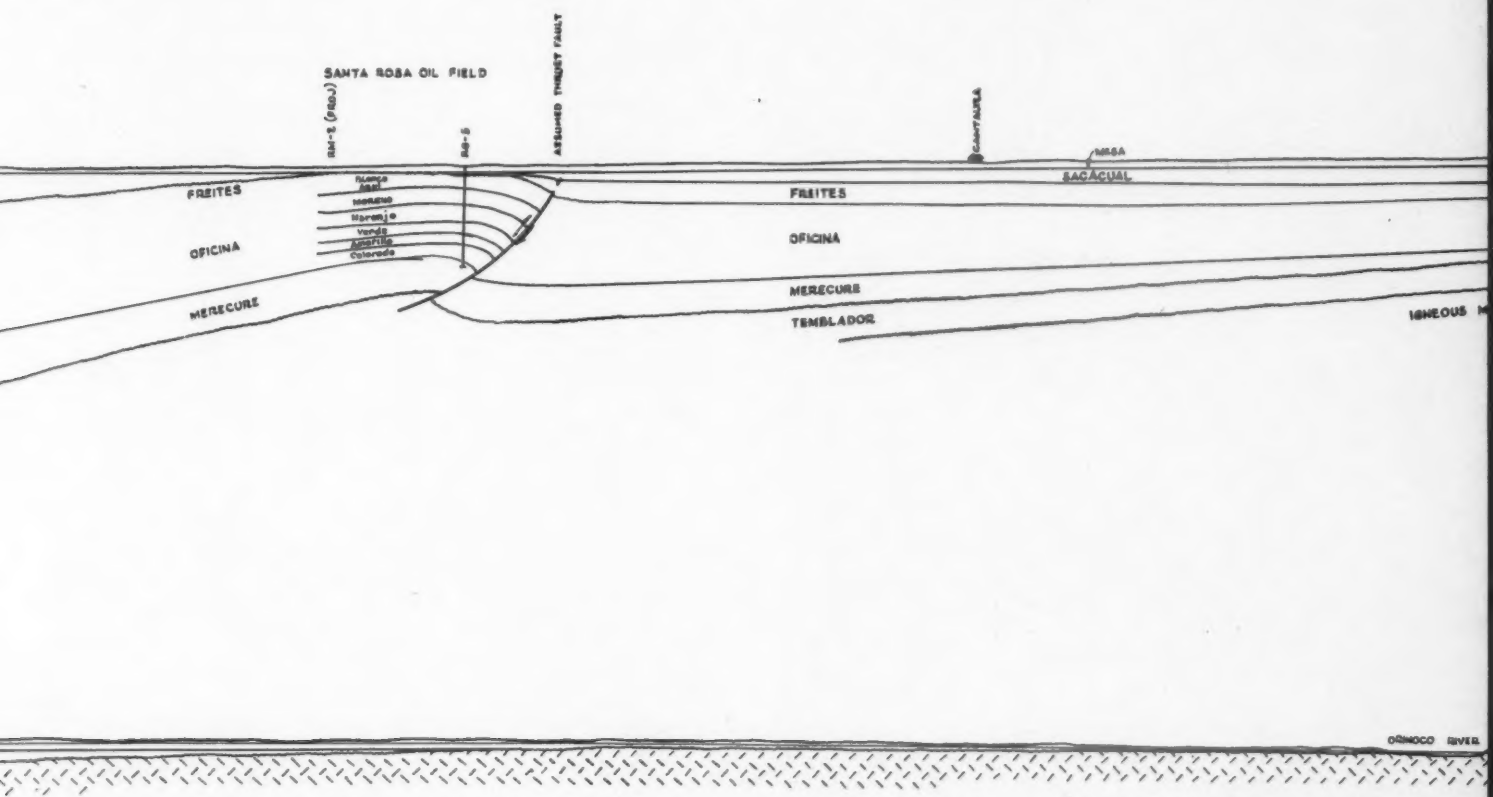


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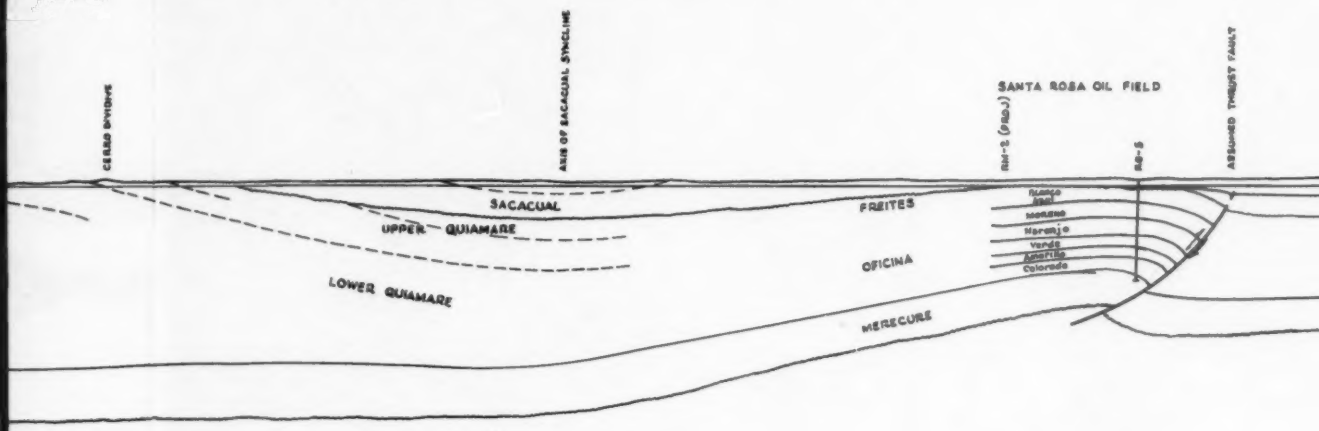


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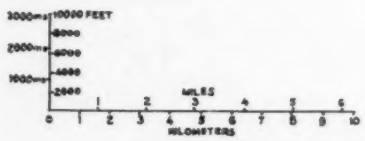


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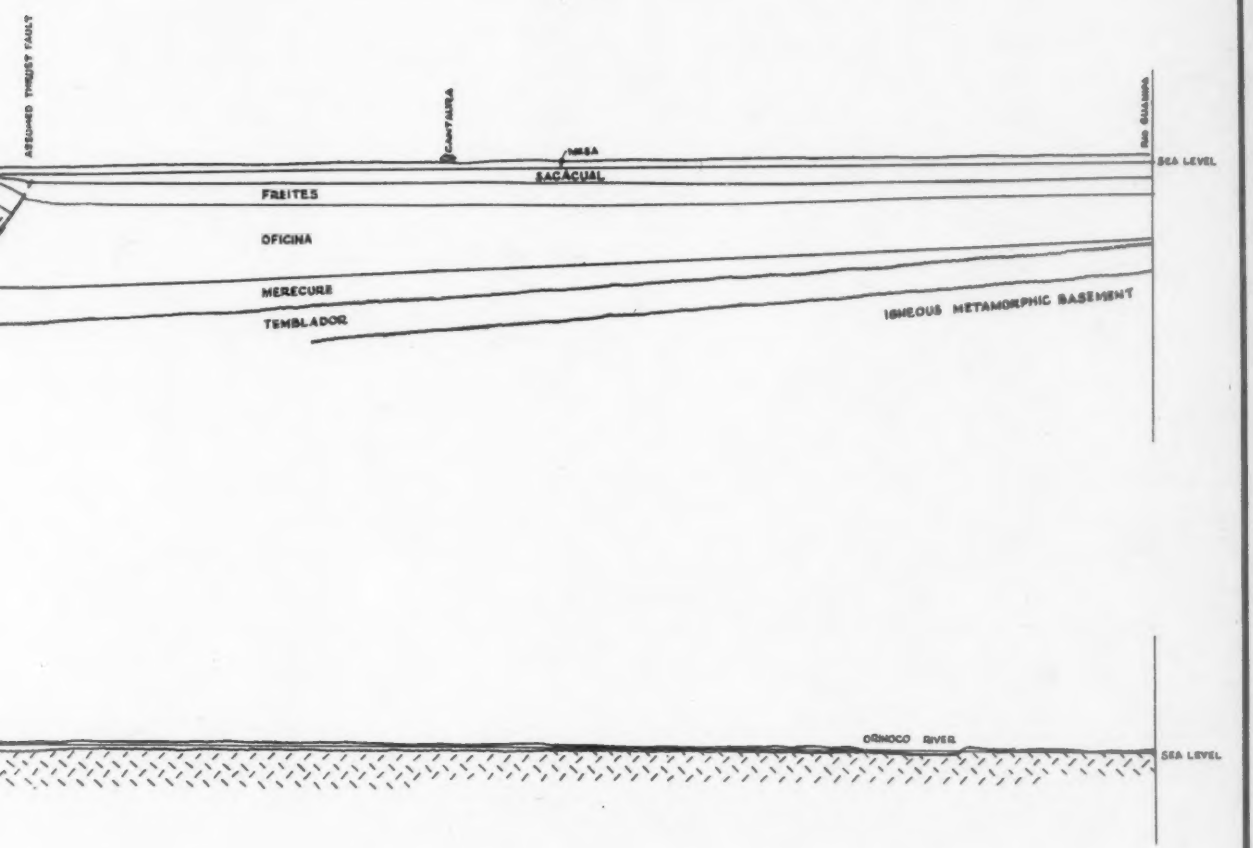
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 61, PP. 1173-1216, 6 FIGS., 11 PLS. NOVEMBER 1950

# GEOLOGY OF THE EASTERN VENEZUELA BASIN (ANZOATEGUI-MONAGAS-SUCRE-EASTERN GUARICO PORTION)

BY HOLLIS D. HEDBERG

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## Abstract

The Eastern Venezuela Basin is a depression in the igneous-metamorphic rock basement filled with Mesozoic and Cenozoic sediments to a thickness of as much as 40,000 feet. The basin has a general east-west trend and is some 800 kilometers long and 250 kilometers broad. It is asymmetrical in cross section with a long gentle southern limb flanking the continental nucleus of the Guayana Shield and an abruptly rising northern limb formed by the folded and faulted Caribbean Ranges. The basin has been the site of geosynclinal deposition through Mesozoic and Cenozoic time with the axis of deposition shifting progressively southward and nearer to the continental shield with successive orogenic episodes in the northern borderland.

The region may be divided into physiographic provinces which in general reflect the areal distribution of formations and the regional geologic structure. These are the Northeastern Mountain, Northwestern Mountain, Foothills, Coastal Plains, Delta, Mesa (and Dissected Mesa), and Rolling Plains provinces.

The basin is bordered on the south by igneous and metamorphic rocks of the Guayana Shield and on the north by the mildly metamorphosed sediments of the peninsulas of Paria and Araya. No igneous or metamorphic rocks are known within the interior of the basin except where deep wells have penetrated the basement.

The oldest sediments are the Hato Viejo and Carrizal formations known only through wells in the southern portion of the basin. These formations are poorly dated but are tentatively considered of early Mesozoic age. They are unconformably overlain by neritic and nonmarine clastics of the Cretaceous Temblador formation.

At the northern edge of the basin the oldest sediments are the formations of the Lower Cretaceous Sucre group (5000–6000 feet) which consists of nonmarine clastics at the base grading upward into reef limestones and other shallow-water marine sediments. The Sucre group is overlain conformably by the Upper Cretaceous Guayuta group (3000–4000 feet) of carbonaceous-bituminous limestones, shales, and cherts representing a deeper-water environment with poor bottom circulation. These in turn give way to the Santa Anita group (3000 feet) of Late Cretaceous-Tertiary age consisting largely of shallow-water marine sandstones, calcareous-dolomitic siltstones, and glauconite beds.

Epeirogenic movements in early Eocene time resulted in widespread emergence. The Upper Eocene-Oligocene Merecure group of sandstones, orbitoid limestones, and coal-basin sediments (6000–7000 feet) marks the beginning of a new cycle of deposits, unconformable on the Cretaceous sediments except in the deepest portion of the geosyncline. The Merecure group is known to extend to both flanks of the basin. It is followed by the even more widespread marine and brackish-water deposits of the Oligocene-Miocene Santa Inés group which locally is 20,000–30,000 feet thick. The succeeding Miocene-Pliocene Sacacual group (as much as 5000 feet) marks a return to brackish-water or nonmarine deposition. At the northern edge of the basin it is markedly unconformable on the Santa Inés. The Pleistocene Mesa formation forms a blanket of several hundred feet of continental deposits, laid down following the last major uplift of the northern mountains and extending over almost the whole area of the basin.

Each of the numerous individual formations comprising the above-mentioned groups is described briefly with respect to name, type locality, areal distribution, lithology, thickness, surface expression, stratigraphic relations, environment of deposition, paleontology, age, and correlation, using both outcrop and well data available throughout the basin. The geologic history of the basin is outlined.

## Foreword

This report was prepared in 1946 as a contribution to a symposium of some 20 papers on the regional geology of Venezuela scheduled for preparation by various authors to accompany the Geologic-tectonic map of Venezuela compiled by Professor W. H. Bucher of Columbia University under the auspices of the Servicio Técnico de Minería y Geología of the Ministerio de Fomento of Venezuela. The scope of the report was largely limited to that portion of the Eastern Venezuela Basin included in the States of Anzoátegui, southern Sucre, western Monagas, and eastern Guárico (Fig. 1), since the geology of bordering areas had been assigned to other contributors. No attempt was made to discuss structure and tectonics since these were to be covered by Professor Bucher in a separate chapter of the

symposium volume. Likewise discussion of economic geology was largely eliminated. Reference for localities mentioned was to be provided by Professor Bucher's map.

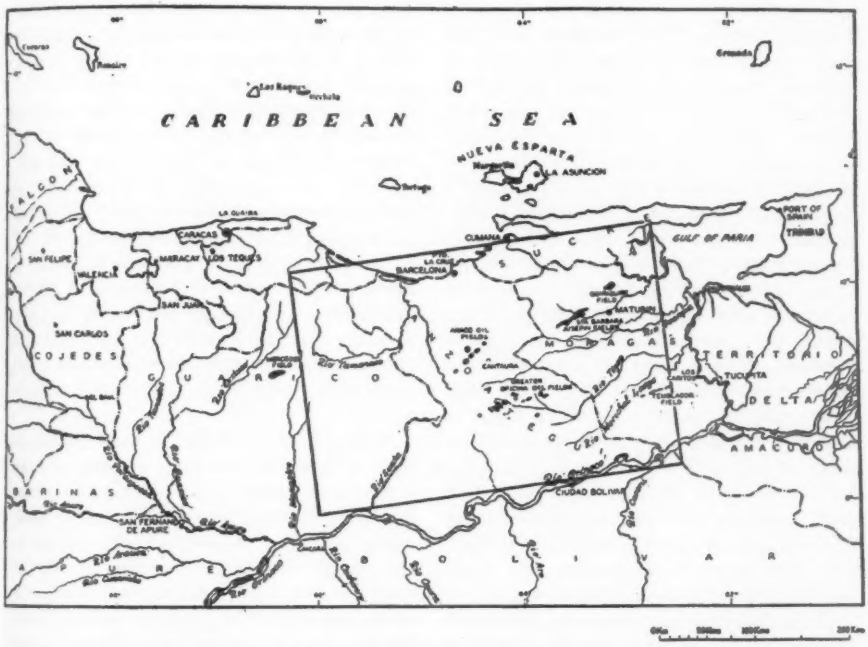


FIGURE 1.—INDEX MAP SHOWING LOCATION IN EASTERN VENEZUELA OF AREA COVERED BY REPORT

For various reasons the symposium project was not completed. However, Professor Bucher's map and contribution on the structure and tectonics of Venezuela are being published by The Geological Society of America (Bucher, 1950; In press), and it appears desirable to also publish such other symposium papers as have been prepared. The writer's original contribution has therefore been edited to bring it more nearly to date and is here presented. Geographic and geologic features not shown on the sketch maps accompanying the text may be found on Professor Bucher's map or on the Boca del Orinoco sheet of the American Geographical Society's 1:1,000,000 map of South America.

Primary acknowledgment is due Dr. Victor López, formerly director of the Servicio Técnico de Minería y Geología of Venezuela, for the conception and motivation of the original symposium project. Only because of his continued and persistent encouragement did the writer take time to prepare the present paper. The report is based primarily upon the author's own investigations, but many data have been derived also from the work of his colleagues in the Mene Grande Oil Company and other organizations to whom appreciative recognition is here given. The original manuscript was read by H. J. Funkhouser, Georges Pardo, Augustin Pyre, and Louis Sass, all of whom made valuable suggestions and criticisms.

## REGIONAL GEOLOGIC SETTING

The term *Eastern Venezuela Basin* is here used to indicate the regional structural basin formed by the contour of the *basement* (*igne-*

*ous-metamorphic*) *rock surface* in Eastern Venezuela as it now exists. The axis of this basin is thus a line marking the present maximum below-sea-level depth to basement. It lies closely adjacent to the northern edge of the basin and extends in an approximately east-west direction for 800 kilometers, from the Orinoco delta westward across the States of Monagas, Anzoátegui, and Guárico to the State of Cojedes (Fig. 4). The trend of this axis is slightly north of east, with a pronounced dog-leg in northern Anzoátegui. The basin does not truly terminate to the west but becomes constricted between the El Baúl igneous-metamorphic mass and the Serranía del Interior. Farther west it changes abruptly to a northeast-southwest trend paralleling the Venezuelan Andes and again opens out to form the *Apure Basin*. The average breadth of the Eastern Venezuela Basin is about 250 kilometers from the Guayana shield on the south to the basement rocks of the Serranía de la Costa of Eastern Venezuela and the Serranía del Interior of Central Venezuela on the north. Its form is distinctly asymmetrical, with a long gentle southern limb and an abruptly rising northern limb (Pl. 1, cross section). The thickness of sediments in the deepest part of the basin is probably more than 40,000 feet.

Of fundamental importance to the understanding of the regional geologic setting of Eastern Venezuela is the distinction between this present *structural* basin and the series of *depositional* basins which existed throughout the Cretaceous and Tertiary. Sedimentation from Cretaceous time onward has been controlled largely by the position of the axis of the so-called *Eastern Venezuela Geosyncline* between the continental mass of the Guayana shield on the south and an actively southward-pushing borderland on the north to which the name Paria has been given.[1] Communication with the open ocean has been largely to the east or northeast. Southward pressure of the mobile Parian borderland and consequent buckling of the northern edge of the sedimentary trough has repeatedly shifted the axis of the geosyncline (maximum thickness of deposition during any one epoch) southward (Fig. 2). Folded mountain and foothill areas have been formed in the northern part of the region from the great thicknesses of sediments marking the axis of the geosyncline during earlier epochs, and the depositional basins of late Tertiary and Quaternary time lie considerably south of the axis of the structural basin as defined by the present attitude of the basement rock surface. The trend of these later depositional basins departs somewhat from the trend of the fundamental structural basin. Thus the N. 80° E. trend of the Late Tertiary depositional basins from the vicinity of Santa Ana to the Pedernales oil field, near the mouth of Caño Manamo (Fig. 2), is relatively superficial and at an angle to the more nearly east-west trend of the sedimentary filling of Eastern Venezuela as a whole.

Another point significant to the understanding and interpretation of the regional geology of Eastern Venezuela is the profound difference in depositional history and character of sediments depending on position with respect to the axis of the geosyncline (Fig. 2). In the axial portions of the series of southward-shifting basins, deposition has been essentially continuous. At the northern edges of these basins active erosion has accompanied the successive orogenic movements of the Parian borderland with the resultant formation of angular unconformities, sharp formational contacts, graywacke-type sandstones, thick conglomerates, and rapid lateral changes from marine to nonmarine deposits. Along the more stable southern limbs of these basins, bordering the Guayana shield, sandstones tend to be more purely quartzose, lithologic units are more persistent, and marked angular divergences of strata are largely lacking though hiatuses, erosional unconformities, and overlaps have resulted from epeirogenic movements. Also, since the geosyncline has almost always opened toward the east or northeast, sediments in general become less marine westward or southwestward in the basin, and unconformities and hiatuses become more pronounced.

[1] The expression "borderland of Paria" is used here to apply to any sort of land area or source of sediments (including "island arcs") which may have lain north of the geosyncline and supplied sediments to it.

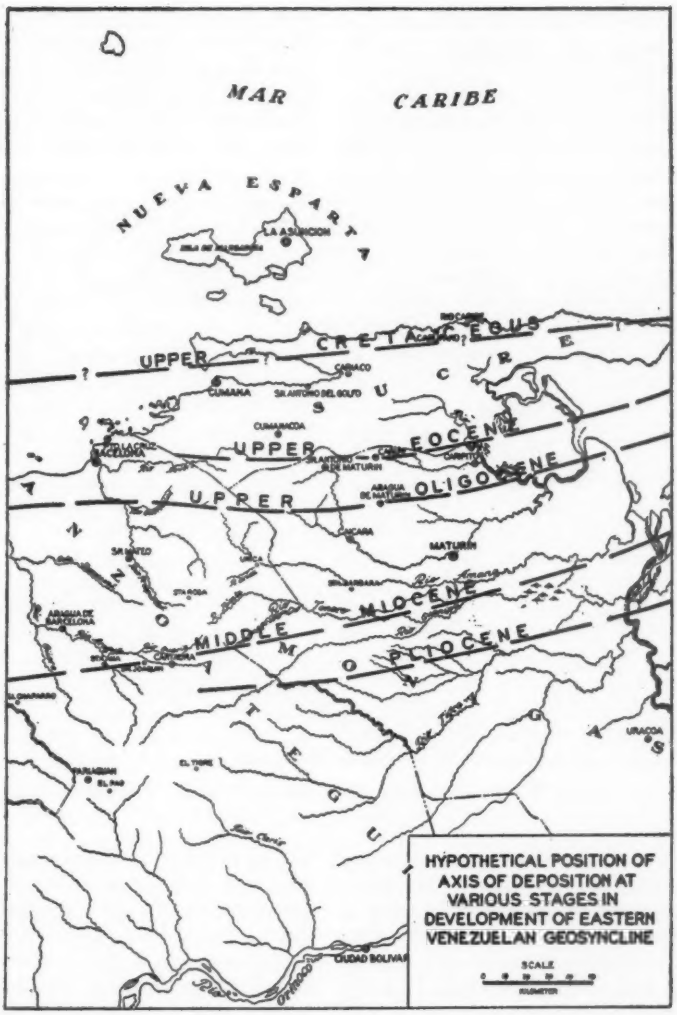


FIGURE 2.—HYPOTHETICAL POSITION OF AXIS OF DEPOSITION AT VARIOUS STAGES IN THE DEVELOPMENT OF THE EASTERN VENEZUELA GEOSYNCLINE

Our picture of the sediments of Eastern Venezuela is based not only on their outcrop character but also on the data derived from numerous deep wells. However, it should be stressed that in the case of the pre-Miocene formations data are largely limited to a relatively narrow east-west belt exposed along the Serranía del Interior which was, at the time of deposition of these sediments, *south* of the axis of the basin. While these older strata have locally been penetrated to some extent in wells within the basin, data regarding their character in a north-south direction across the basin are very inadequate.

## Physiographic Setting

### *Physiographic Provinces*

*General picture.*—The Eastern Venezuela Basin has here been divided into physiographic provinces which in general closely reflect the areal distribution of formations and the geologic structure (Fig. 3).[2] Naturally any large-scale division into provinces must be made in a free-hand manner which, while outlining *dominant* physiographic character, will include many local variations from the type. Likewise, the bases used for drawing province boundaries are to a considerable extent a matter of individual viewpoint and probably no two geologists would draw them the same.

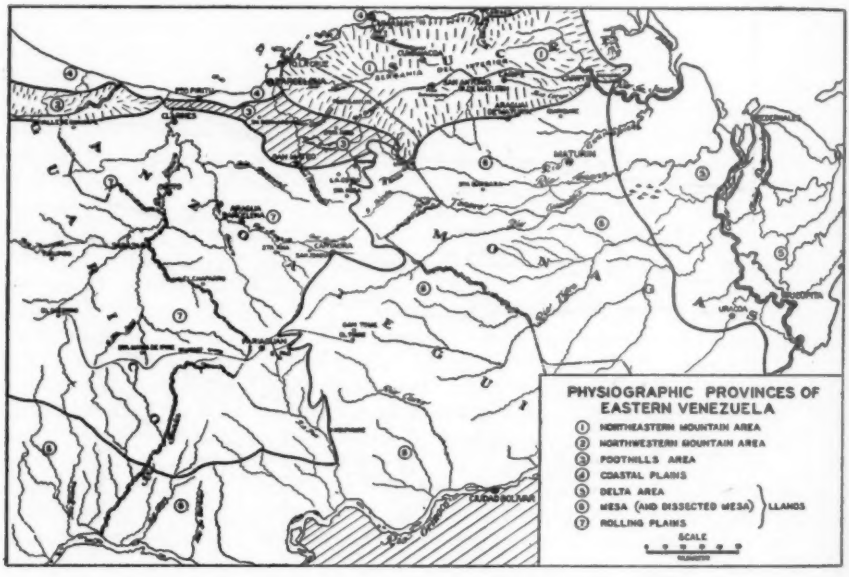


Figure 3.—Physiographic Provinces of Eastern Venezuela

1. *Northeastern Mountain Province* (Fig. 3; Pls. 2, 3).—This province includes that portion of the Serranía del Interior of Venezuela between Barcelona on the west and Caripito, Guanoco, and El Pilar on the east, and between the foothills belt and mesa province on the south and the Cariaco depression on the north. This mountain area is almost entirely folded and faulted Cretaceous sediments, with some Eocene along its souther border. Rugged ridges, peaks, and flat-topped mountains are held up by the thick indurated sandstones and limestones of these ages. Relief varies from sea level at the western end where the islands near Guanta represent the partially submerged end of the range (Pl. 3, fig. 3) to the heights of Pico Turumiquire which rises to about 8500 feet (Pl. 2, fig. 1). The range in general trends slightly north of east, but swings northward at its western end. Drainage to the west is principally by the Río Neverí and its tributaries; to the east by the San Juan and Caripe rivers; and to the north by the Manzanares and Cariaco rivers. To the south numerous streams, with shorter mountain courses, cut across the strike of the frequently vertical and steeply dipping strata of the southern edge of the mountain chain forming deep and difficultly passable gorges in the more resistant beds. The

[2] *Also see* Boca del Orinoco sheet of American Geographical Society's 1:1,000,000 map of South America.

interior of the mountain area is thoroughly folded and faulted, but dips are often moderate, giving rise to large tilted blocks capped by resistant sandstones or limestones and to broad anticlinal valleys in the softer strata. The thick upper Cretaceous shaly limestones and shales form a very distinctive topography of sharp V-shaped ridges and valleys contrasting with the blocky topography of the more resistant sandstones of both younger and older units.

2. *Northwestern Mountain Province* (Fig. 3).—The northern edge of the Eastern Venezuela Basin west of the Río Unare is formed by the Serranía del Interior of Central Venezuela. Here the mountain area of Cretaceous and early Tertiary sedimentary rocks is more limited in breadth than in the area east of Barcelona. Thrust faulting is prominent, and there is a very distinct alignment of ridges parallel to the general N. 80° E. trend of the range. Drainage is largely to the south. A series of limestone "morros" of Paleocene (?) age appears at intervals along this mountain front and furnishes low but distinctive topographic features (Pl. 3, fig. 2).

3. *Foothills Province* (Fig. 3).—At the southwestern edge of the Northeastern Mountain Province and extending across the gap between this province and the Northwestern Mountain Province is a more or less continuous belt of foothills formed largely in Eocene, Oligocene, and Miocene sediments. These foothills consist largely of low hogback strike ridges of sandstone and conglomerate which frequently outline folded structures. Particularly prominent are the San Bernardino conglomerate hills forming the principal relief in the Barcelona gap. Drainage in the foothills belt is generally of the block or trellis type with numerous antecedent streams cutting across resistant strike ridges. Many of the streams are intermittent; the water derived from mountain sources being lost in the foothill belt. Both east and west of the area outlined on the accompanying map (Fig. 3) the foothills belt is greatly reduced, and the mountains rise rather abruptly from rolling plains. However, still farther west in the State of Guárico west of Altagracia de Orituco a broad foothills belt is constituted by the so-called "galeras"—a succession of prominent sandstone strike ridges.

4. *Coastal Plains Province* (Fig. 3).—Along the shore of the Caribbean Sea westward from Barcelona to Higuerote is a narrow strip of flat coastal plain representing Recent or Pleistocene sea floor (Pl. 3, fig. 3). This plain is also locally developed along the northern edge of the Northeastern Mountain Province, between Barcelona and Cumaná, and forms the Cariaco depression between the Serranía del Interior and the Serranía de la Costa, east of the town of Cariaco. There are many broad lagunas and salt marshes in this coastal belt, and during the rainy season much of the area is continually under water. Deposits of recent marine shells at the inner edge of the belt only a few feet above sea level, as well as physiographic evidence, indicate that it was recently a part of the coastal sea floor.

5. *Delta Province* (Fig. 3).—The easternmost part of the Eastern Venezuela Basin consists of a swampy lowland area comprising the deltas of the Orinoco, Guanipa, San Juan, and other streams flowing eastward into the Gulf of Paria. This area is characterized by a maze of distributaries many of which have been scoured to great depths by tidal action to form so-called *caños*. Tidal effects extend for many kilometers, inland along these channels. For example, at Caripito on Rio San Juan the daily rise and fall of the tide is through a range of 13–14 feet. Broad intertidal mudflats exist along much of the coast. Part of the delta area is heavily jungled, but there are numerous broad open savannas. This province is described in some detail by Liddle (1928).

6. *Mesa (and Dissected Mesa) Province* (Fig. 3; Pl. 4).—The mesa region of Eastern Venezuela constitutes one of the most distinctive physiographic units of the country. The Venezuelan mesas are not the small isolated flat-topped hills for which the term is commonly used in Western United States, but are broad flat to gently rolling, grass-covered plains, usually bounded by low but precipitous cliffs which separate them from the adjacent lower country. They are typically formed in the Mesa formation of Pleistocene age and owe their distinctive appearance to the lithologic character of this formation which consists of relatively unconsolidated porous sandstones, siltstones, and sandy claystones covered by a hard pebbly ferruginous capping.

The mesa surface as a whole slopes gently

eastward and southward. The eroded western edge is marked by an "almost continuous line of vertical cliffs (*farallones*) extending from the northern mountain front southwestward across central Anzoátegui nearly to the Orinoco River. A large southeastward-projecting embayment into the Mesa Province just north of Cantaura extends to and beyond the Guanipa River. Another similar embayment extends southeastward from Pariaguan along the course of the Pao River. In southern Anzoátegui the mesa front turns westward 20–50 kilometers north of the Orinoco River. It then trends northwestward through Guárico leaving a continually widening belt of mesa north of the Orinoco and Apure rivers. The elevation of the mesa surface along its western front in Anzoátegui ranges from about 650 feet above sea level at Atapirire in the south to more than 1200 feet near La Ceiba to the north. Southward the mesa surface extends even beyond the Orinoco River, while eastward it merges gradually with the lowlands of the Orinoco delta.

The mesa surface is in a youthful physiographic stage. Drainage is poor, and many areas are characterized by circular undrained depressions reaching several hundred meters in diameter (Pl. 4, fig. 1). Streams flow eastward and southward and characteristically start with spring-fed marshy water courses called *morichales* from the characteristic growth of "moriche" palms (*Mauritia* sp.) which accompanies them (Pl. 5, fig. 1). The mesa streams are typically perennial as contrasted with the intermittent and seasonal character of streams of the adjacent rolling plains to the west. The typical vegetation of the mesa surface is coarse grass with sparsely scattered *chaparro* trees and with local concentrations of *moriche* palms and other trees along stream courses (Pl. 4, fig. 2).

The mesa is not one continuous plain but is actually a series of plains separated by narrow steep-walled elongate valleys where major streams have cut through the Mesa formation. Locally in the vicinity of these streams there are large areas of partially dissected mesa where a typical badlands topography is developed (Pl. 5, fig. 2). Also about the borders of the mesa front there are frequently large areas of secondary mesa or mesa outwash. Indications of more than one mesa cycle also exist in the occasional flat-topped hills rising above the present plains (Pl. 4, fig. 2). In Anzoátegui and Monagas the mesa capping tends to slope gently toward the major eastward-flowing stream courses and tends to be gently arched across the interstream areas.

7. *Rolling Plains Province* (Fig. 3).—The great interior plains of Eastern Venezuela between the Serranía del Interior on the north and the hills and mountains of the Guayana shield on the south are generally known as the Venezuelan *llanos*. This term includes not only the Mesa Province and at least part of the Delta Province discussed above, but also the broad region of moderately dissected and moderately wooded rolling plains in Anzoátegui, Guárico, and other states west of the main mesa front. To differentiate this portion of the *llanos* from the Mesa Province it is here referred to as the *Rolling Plains Province*. This province is developed almost entirely in low-dipping, moderately soft, shaly Miocene and Pliocene sediments. Relief in any one area is usually no more than 100–200 feet. The northern part of the province is dominantly wooded or brush-covered, whereas the southern part is semi-savanna country with wooded areas largely confined to stream courses. (Where clay soils are particularly predominant large areas are characteristically covered with a scrubby palm, "Palma llanera" (*Copernicia tectorum* Mart.) (Pl. 4, fig. 3). Erosion has reached a mature stage, contrasting with the youthful stage of the mesa surface. Streams are largely intermittent.

### *Drainage*

(*See* Figs. 1 and 3)

The present drainage of Eastern Venezuela is less dominated by the Río Orinoco than might be expected considering the overwhelming size of that stream compared to any others. The western part of the basin is indeed drained almost entirely by tributaries of the Orinoco (Manapire and Guariquito) or of its major tributary, the Apure (Guárico and Portuguesa), but east of the Río Manapire the southward-flowing tributaries of the Orinoco proper are relatively short, and drainage is either to the

north through the Unare and Neverí systems flowing into the Caribbean Sea or to the east through the Tigre, Guanipa, Amana, and San Juan river systems which either debouch into the delta water courses or empty directly into the Gulf of Paria.

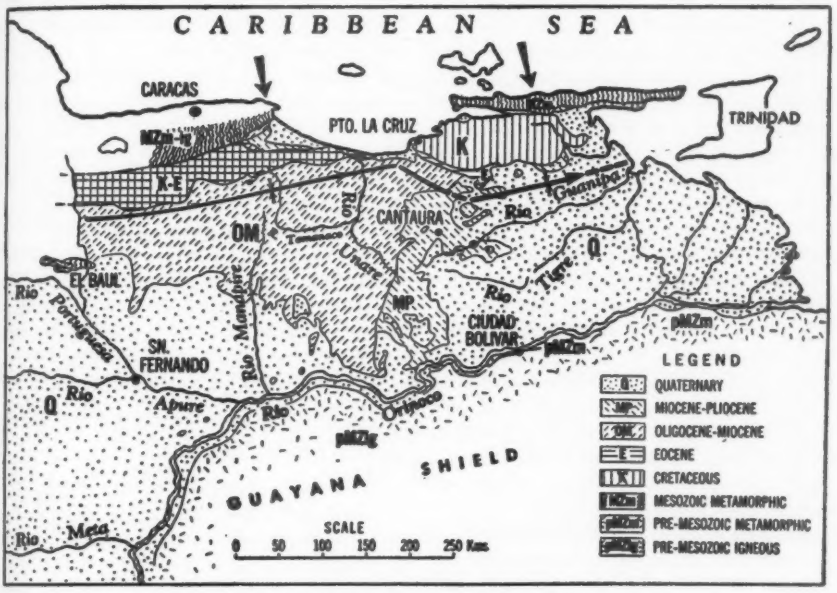


FIGURE 4.—AREAL GEOLOGIC SKETCH MAP

The Unare and Neverí systems with their direct outlet to the Caribbean Sea have evidently captured a considerable portion of the mountain drainage which originally escaped southward to the Orinoco or eastward to the delta region. An outstanding example of such piracy is the Río Tamanaco which, starting less than 20 kilometers from the Caribbean coast in the mountains of northeastern Guárico, flows southward for 35 kilometers, then abruptly turns eastward, and finally exits to the Caribbean Sea through the Río Unare. Likewise the Río Orégano flowing southward from the Serranía del Interior east of Santa Inés is abruptly diverted westward to join the Aragua-Neverí system leaving a dry channel in the mesa between Piripire and Urica where it once probably connected with the eastward-flowing Río Tácata. Still farther east there is good evidence in a broad valley and line of lagoons on the mesa surface south of Urica that the Amana River has beheaded a large mountain stream that once flowed southward as far as the Tonoro River.

The Barcelona gap, the Unare drainage system, and the sharp bend in the Orinoco at Caicara have suggested to many that the Río Orinoco may originally have flowed northward into the Caribbean Sea and that its present eastward course beyond Caicara may be due to capture by streams emptying into the Atlantic. However, there is little supportive evidence. Rather, it would seem that the Unare system is the younger and that it has been growing at the expense of the Orinoco.

## IGNEOUS AND METAMORPHIC ROCKS

(Fig. 4)

Exposures of crystalline rocks are lacking in the interior of the Eastern Venezuela Basin. However, igneous and metamorphic rocks of the Guayana shield border the basin on the south, and on the north are the metamorphosed sediments of Paria and Araya (Serranía de la Costa) and the igneous and metamorphic rocks

of the Serranía del Interior of Central Venezuela. (No occurrences of either igneous or metamorphic rock are known in the Serranía del Interior of Eastern Venezuela.) At the western end of the Eastern Venezuela Basin the igneous-metamorphic mass of El Baúl with smaller associated igneous rock exposures constitutes a partial barrier between this basin and the Apure basin farther west. The only known occurrences of igneous rock intruding the sedimentary filling of the basin are in the northwestern part of the region in the general vicinity of San Juan de los Morros where basic igneous material has been complexly injected into Eocene and Cretaceous sediments.

The southern border of the Eastern Venezuelan basin approximately coincides with the course of the Orinoco River. While Pleistocene Mesa formation deposits are quite widespread south of the river, none of the other formations of the Cretaceous-Tertiary geosyncline are known there, and the river channel is largely carved in the basement rocks of the Guayana shield. A few inliers of this basement rock are exposed in and just north of the river—largely granites and granitic gneisses although there are some other rock types. At La Peña-Soledad on the north side of the river opposite Ciudad Bolívar is some coarse-grained gabbroid rock. Cerro Cabruta, just north of Caicara, is a large mass of biotite-amphibole granite extending some 12 kilometers northeastward from the Orinícо River (Bendrat, 1911). Cerro Iguana (Lomas Parmanitas) 10 kilometers north of the Orinoco in southeastern Guárico is quartzite (some of which is ferruginous). Cerro Las Peñas, 40 kilometers farther west and 15 kilometers north of the Orinoco, is gneissic granite. The ferruginous quartzite of the iron-producing Imataca series outcrops just south of the river east of Ciudad Bolívar but is not exposed on the north side.

Of particular significance with respect to the composition of the basement rocks underlying the Eastern Venezuela sedimentary basin are the data from some 20 petroleum exploration wells in southern Monagas, Anzoátegui, and Guárico. These wells have encountered basement at depths of 3000 to 6000 feet below sea level and show a wide variety of basement rock types, including biotite granite, schist, granitoid gneiss, granodiorite, quartz diorite, altered greenstone, and quartzitic phyllite. Granitic and dioritic igneous rocks and crystalline schists are the dominant types. Most of the rocks show gneissic or schistose structure with steep to vertical dips. Geophysical work indicates a strike of about N. 70° E. for the structure of the basement rock through much of this southern area. This is comparable to the common trend of lineation in the exposed portion of the Guayana shield south of the river. The strike of the basement *surface* as deduced from its elevation in the various wells varies from northeast in southern Monagas to east-west in southern Anzoátegui, and slightly north of west in southern Guárico. The dip of this surface is 1°–2° N., steepening northward. The basement surface is usually weathered to a depth of less than 10 feet though in one well weathering was noted to a depth of 40 feet.

## General Statement on Sedimentary Rock Units

In succeeding sections of this report the principal sedimentary rock units (groups, formations, members) of this part of Eastern Venezuela are described in order from oldest to youngest. The description of each unit usually covers origin of name, type locality, character of type section, areal distribution, general lithology, thickness, surface expression, stratigraphic boundaries and relations to other units, environment of deposition, age, and correlation. Information from both well and outcrop sections has been used. To keep the descriptions reasonably brief it has been necessary to limit severely the detailed paleontologic, mineralogic, and lithologic data. For areal distribution reference should be made to the geologic sketch map of Figure 4 and to Bucher's Geologic-tectonic map (Bucher, 1950). Generalized correlation of formations (1) along the northern edge of the basin, (2) across the basin from north to south, and (3) along the southern edge of the basin is shown in Figure 5.

## Triassic-Jurassic (?) Strata

### *Hato Viejo and Carrizal Formations*

The oldest sedimentary rocks known in the Eastern Venezuela Basin belong to the Hato

FIGURE 5.—CORRELATION CHART OF FORMATIONS IN EASTERN VENEZUELA

Viejo and Carrizal formations penetrated by a number of wells in southwestern Anzoátegui and in Guárico. These formations unconformably underlie the Cretaceous Temblador formation and have been considered tentatively to be of early Mesozoic age (Hedberg, 1942). The following description of the type section of the Hato Viejo formation was made by M. W. Haas of Creole Petroleum Corporation, November 24, 1939:

HATO VIEJO FORMATION: "Named for Hato Viejo No. 1 well drilled by the Standard Oil Company of Venezuela in the District of Monagas, State of Anzoátegui. In this well the formation was entirely cored showing its relation to the argillite (Carrizal formation) and the basement. The type section is Hato Viejo No. 1 from 2740′ to a depth of 2850′. The formation directly underlies the argillite (Carrizal formation) and overlies the basement. Its contact with the argillite (Carrizal formation) appears to be conformable.

"Essentially the formation consists of massive hard fine to coarse grained friable gritty pinkish gray, gray and dark gray sandstone which is slightly calcareous, and in portions is extremely micaceous and pyritic. The grains are rounded and firmly cemented. Quartzite cobbles and conglomerates occur in the sandstone at irregular intervals and jointing is noted in several places. Calcite veins are present along fracture planes. Certain portions of the sandstone show quartzitic characteristics and others are highly feldspathic, green, and glauconitic."

The type section of the Carrizal formation was designated as follows by J. A. Tong of the Socony-Vacuum Oil Company, January 5, 1940:

"The Carrizal formation is defined as those hard argillite or silty rocks (in the southern part of the Venezuelan Llanos area) underlying the Temblador or Oficina formations and overlying a sandstone formation designated as the Hato Viejo formation. In some cases it may lie directly on the basement complex. The upper part of the Carrizal formation is typical in Socony-Vacuum Oil Company's Carrizal No. 1 well from 3645′ to bottom at 5243′. The lower part is typical in Socony-Vacuum Oil Company's Suata No. 1 well between 1503′ and 2550′. The upper part of the latter may be the equivalent of some of the lower part of the former, but because of the lack of markers the amount of overlap, if any, is not known. The upper contact is marked in some cases by oxidation coloring, possibly due to weathering on the old erosion surface. The lower part is marked by argillite beds containing pebbles and conglomerate beds."

The descriptions of the type sections of these two formations may be elaborated somewhat on the basis of data from other wells. The Hato Viejo formation is characteristically medium- to coarse-grained, variably arkosic and quartzitic sandstone which grades upward into the Carrizal formation. It is locally at least 300 feet thick and may be considerably thicker. The lower part of the unit carries a simple zircon-leucoxene suite of minerals which, however, grades upward into a garnet-biotite suite like that of the overlying Carrizal formation. Some streaks of green argillite in the Hato Viejo formation also indicate its close relation to the Carrizal formation. No fossils have been found.

The Carrizal formation is very distinctive lithologically, consisting typically of tough massive dense greenish argillite. It is remarkably homogeneous, although variably silty and locally possessing intercalated beds of sandstone and even pebble conglomerate. It is generally noncalcareous, but some thin siltstone laminae as well as some vertical joints are cemented with calcite. All samples give a strong reaction for phosphate. The formation is known to be at least 2000 feet thick and is probably much thicker in full development. The only fossils obtained to date are fish remains (teeth, scales, and bones), some poor indeterminate arenaceous Foraminifera, and a questionable linguloid brachiopod. Mineralogically the formation is characterized by garnet and biotite. Feldspars are common.

An angular unconformity separates the Carrizal and Hato Viejo formations from the overlying Cretaceous Temblador formation, and the upper surface of the Carrizal formation is deeply weathered. Moreover, the Carrizal-Hato Viejo sediments are much more highly lithified than is the overlying Cretaceous. Little direct evidence is available regarding the age of these formations. However, the writer (1942, p. 201) has indicated some reasons for supposing them to be time equivalents of the Jurassic-Triassic La Quinta formation of the Venezuelan Andes. More recently drilling in central Guárico has further supported this idea by showing beds with typical red La Quinta lithology occupying a position similar to that of the Carrizal formation below the Cretaceous Temblador formation. However, the Carrizal-Hato Viejo beds may even be Paleozoic, and certainly further evidence is necessary before any posi-

FIGURE 1. CRESTAL PORTION OF SERRANÍA DEL INTERIOR, STATES OF SUCRE AND MONAGAS
Pico Turumiquire in central background is the highest point in northeastern Venezuela and reaches an elevation of about 8500 feet.

FIGURE 2. INTERMONTANE VALLEY IN CRETACEOUS SEDIMENTS, MUNDO NUEVO, MONAGAS

FIGURE 3. GORGE OF RÍO GUAYUTA IN CRETACEOUS SEDIMENTS AT MOUNTAIN FRONT EAST OF ARAGUA DE MATURÍN, MONAGAS

PHYSIOGRAPHY OF NORTHEASTERN MOUNTAIN PROVINCE

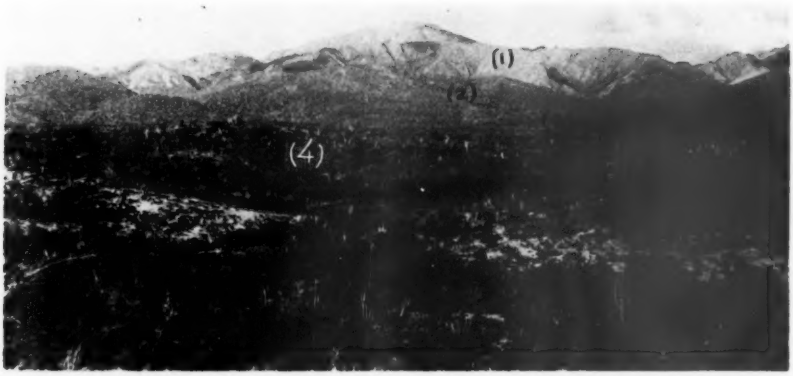


FIGURE 1. SOUTHERN FRONT OF SERRANÍA DEL INTERIOR NEAR SANTA INÉS, ANZOÁTEGUI
Río Querecual valley at left. (1) Cretaceous, (2) Eocene, (3) Oligocene, (4) Miocene.

FIGURE 2. LIMESTONE *morros*, SIMILAR TO THESE AT SAN JUAN, CONSTITUTE DISTINCTIVE TOPOGRAPHIC FEATURES IN MANY PLACES ALONG THE GÜARICO MOUNTAIN FRONT

FIGURE 3. COASTAL PLAIN NEAR PUERTO LA CRUZ, ANZOÁTEGUI
Note mangrove fringe along shore. In background western end of Serranía del Interior plunges into Caribbean Sea.

MOUNTAIN FRONT OF SERRANÍA DEL INTERIOR. COASTAL PLAIN

FIGURE 1. GRASS-COVERED MESA PLAIN NEAR MATURÍN, MONAGAS
Note characteristic circular undrained depression with scattered trees along border.

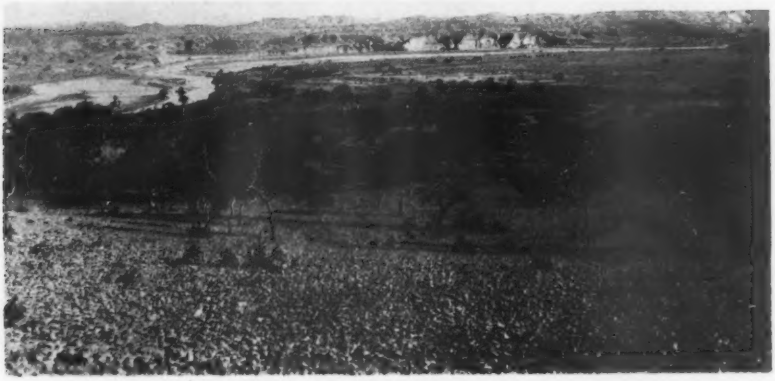

FIGURE 2. TĀCATA RIVER IN NORTHERN ANZOÁTEGUI
Chaparro trees, typical of the Eastern Venezuelan mesa, in foreground. Moriche palms farther back near river. *Farallones* (cliffs) developed along river and also on upland divide in background.

FIGURE 3. WITHIN THE ROLLING PLAINS PROVINCE
Flat areas are frequently developed on the Freites shale outcrop and are characteristically covered with the scrubby "*palma llanera*."

LLANOS OF EASTERN VENEZUELA

FIGURE 1. INCIPIENT STREAM COURSE ON TOP OF THE MESA NEAR SAN TOMÉ, ANZOÁTEGUI
Note the cluster of *moriche* palms (*morichal*) which commonly characterizes the beginning of stream courses on the mesa.

FIGURE 2. DISSECTED MESA ALONG MODERATELY DEVELOPED STREAM COURSE NORTHWEST OF SAN TOME, ANZOÁTEGUI

FIGURE 3. GUARAPICHE RIVER IN DISSECTED MESA AREA NORTHWEST OF MATURÍN, MONAGAS

FIGURE 1. TYPICAL CROSS-BEDDED, COARSE-GRAINED BARRANQUÍN SANDSTONE ALTERNATING WITH THIN-BEDDED SHALY SANDSTONE, RÍO POTRERO, NEAR PUNCERES, MONAGAS

FIGURE 2. EL CANTIL, CHIMANA, AND QUERECUAL FORMATIONS FROM LEFT TO RIGHT AT WEST END OF LA BORRACHA ISLAND NORTHWEST OF PUERTO LA CRUZ, ANZOÁTEGUI
Note that whole section is overturned.

FIGURE 3. CHARACTERISTIC WEATHERING OF EL CANTIL REEF LIMESTONE, LA BORRACHA ISLAND, NORTHWEST OF PUERTO LA CRUZ, ANZOÁTEGUI

FORMATIONS OF LOWER CRETACEOUS SUCRE GROUP

FIGURE 1. LARGE DISCOIDAL CONCRETIONS IN BASAL PART OF QUERECUAL FORMATION ON LA BORRACHA ISLAND, NORTHWEST OF PUERTO LA CRUZ, ANZOÁTEGUI

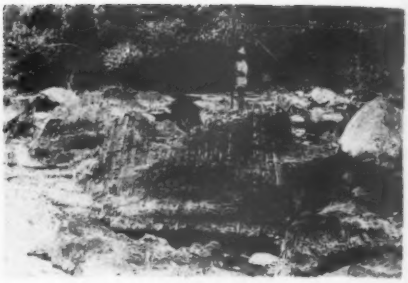

FIGURE 2. THIN-BEDDED CHERTY LIMESTONES OF SAN ANTONIO FORMATION ON RÍO QUERECUAL, ANZOÁTEGUI

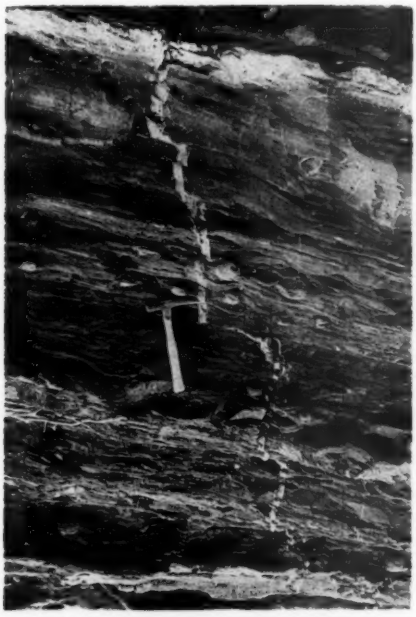

FIGURE 3. FAULTED SANDSTONE DIKE IN GUAYUTA SEDIMENTS, RÍO GUAYUTA, MONAGAS
Dike displaced along bedding planes.

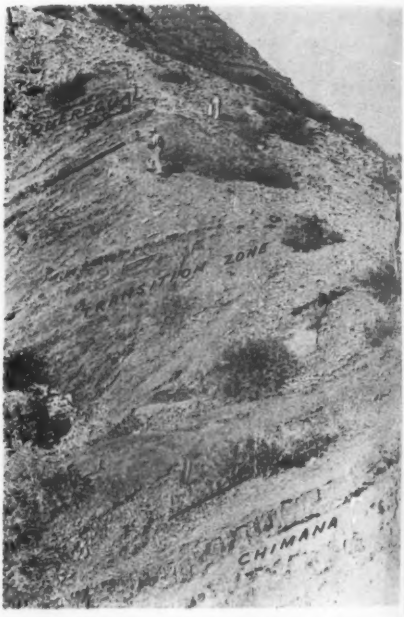


FIGURE 4. CONTACT BETWEEN CHIMANA AND QUERECUAL FORMATIONS AT WEST END OF CHIMANA GRANDE ISLAND, NORTH OF PUERTO LA CRUZ, ANZOÁTEGUI

FORMATIONS OF UPPER CRETACEOUS
GUAYUTA GROUP

FIGURE 1. CHARACTERISTIC ROUNDED SURFACES OF VERY HARD GLAUCONITIC CALCAREOUS AND DOLOMITIC SILTSTONES OF CARATAS FORMATION ON RÍO QUERECUAL, ANZOÁTEGUI

FIGURE 2. *Venericardia* LIMESTONE IN CARATAS FORMATION NEAR URICA, ANZOÁTEGUI

FIGURE 3. WEATHERED AND PITTED CALCAREOUS SILTSTONES OF CARATAS FORMATION IN CONTACT WITH SO-CALLED "BOULDER BED" OF BASAL MERECURE GROUP
At Kilometer 12 on Puerto la Cruz-Oficina highway, Anzoátegui.

UPPER CRETACEOUS-TERTIARY SANTA ANITA GROUP

FIGURE 1. ROUNDED MASSES OF ORBITOID LIMESTONE (SO-CALLED "BOULDER BED") BELONGING TO THE TINAJITAS FORMATION OF THE MERECURE GROUP
At contact of this formation with the Caratas formation at Kilometer 11 on the Puerto la Cruz-Oficina Road, Anzoátegui.

FIGURE 2. THIN BED OF PEBBLY LIMESTONE WITH *Algae* AND *Tubulostium*
At contact between Merecure sandstone above and Caratas siltstone below at El Salto on Río Areo, Monagas. Figures are standing on limestone bed.

FIGURE 3. QUARTZITIC LOS JABILLOS SANDSTONE IN TYPE SECTION ON RÍO QUERECUAL, ANZOÁTEGUI

UPPER EOCENE-OLIGOCENE MERECURE GROUP

FIGURE 1. SANDSTONES AT TOP OF CAPAYA TONGUE ON QUEBRADA CARAPITA, ANZOÁTEGUI

FIGURE 2. CARAPITA SHALE, QUEBRADA CARAPITA, ANZOÁTEGUI

FIGURE 3. CLAY-SHALES AND LIGNITE IN SUATA FORMATION ON RÍO GUERE NORTHWEST OF ARAGUA DE BARCELONA, ANZOÁTEGUI

OLIGOCENE-MIOCENE SANTA INÉS GROUP

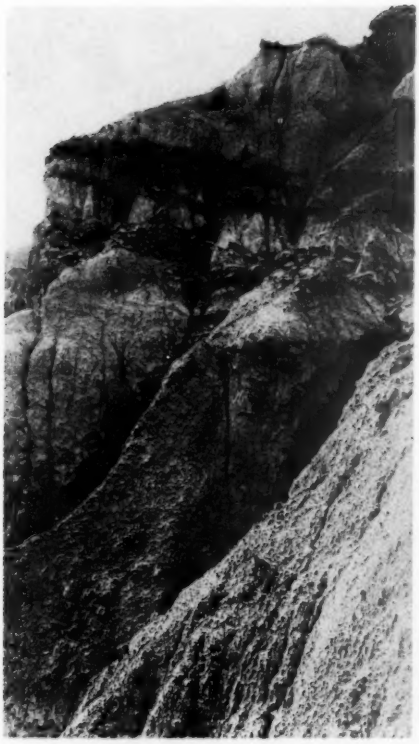

FIGURE 1. UNCONFORMABLE CONTACT BETWEEN GENTLY DIPPING QUIRIQUIRE FORMATION OF SACACUAL GROUP AND VERTICAL CARAPITA SHALE OF SANTA INÉS GROUP
On east slope of Cerro Corozal, near Río Oregano, Anzoátegui.

FIGURE 2. CONGLOMERATES AND CLAYSTONES OF MESA FORMATION NEAR URICA, ANZOÁTEGUI

FIGURE 3. CHARACTERISTIC WALLS (*farallones*) FORMED BY MESA FORMATION SANDSTONE ALONG MESA FRONT
Río Güico near San Tomé, Anzoátegui.

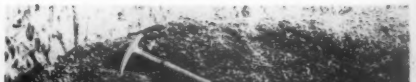

FIGURE 4. MESA CONGLOMERATE IN RÍO GUANIPA VALLEY, ANZOÁTEGUI

PLIOCENE SACACUAL GROUP AND
PLEISTOCENE MESA FORMATION

tive age assignment other than pre-Cretaceous can be made.

Neither La Quinta nor Carrizal-Hato Viejo beds have been definitely identified in outcrop in any portion of the Eastern Venezuela Basin. However, the writer believes that probably at least some portion of the metamorphosed sediments composing the so-called Caribbean series flanking the Eastern Venezuela Basin on the north of the Paria-Araya peninsulas and in the Serranía del Interior of Central Venezuela will be found to be equivalent in age to these formations (Hedberg, 1942, p. 202–203).[3]

[3] Some observers have been impressed by certain lithologic similarities between the Paria-Araya metamorphics exposed on the Carúpano-El Pilar road and the normal Upper Cretaceous sedimentary section and have suggested an Upper Cretaceous age for the metamorphics. The principal point of similarity appears to be the local presence of black chert and lenticular limestone concretions suggestive of the Guayuta group. However, the great bulk of the Carúpano-El Pilar cross section appears to consist of a phyllite (usually noncalcareous) which it is very difficult to picture as the metamorphosed equivalent of the highly calcareous Cretaceous section.

In Central Venezuela in the Barquisimeto area the Cretaceous sequence grades laterally into metamorphic rocks, and Cretaceous sediments occur within the Northern Range of Trinidad. It therefore would not be improbable to find some Cretaceous included in the Paria-Araya area. However, there is certainly equal probability of finding in this area the metamorphosed equivalents of the thick pre-Cretaceous (early Mesozoic?) section known below the Cretaceous in the Eastern Venezuela basin or the definitely confirmed Jurassic of the Northern Range of Trinidad. Phyllites lithologically similar to those of Paria-Araya are known to be associated with and to underlie the ammonite-dated Jurassic of the Trinidad Northern Range. Certainly the lithologic similarity of the known pre-Cretaceous sediments to the metamorphics is as great or greater than that of the Cretaceous rocks.

The general succession from south to north in the Caribbean Cordillera is from younger to older rocks, and it would be less surprising to find a pre-Cretaceous age for the metamorphics of Paria-Araya which are in fault contact on the south with unmetamorphosed Lower Cretaceous sediments than to find these metamorphics to be of Upper Cretaceous age. In the latter case the northern part of the Serranía del Interior would appear to be the crest of a huge geanticline of which Paria-Araya would constitute the northern flank, more highly metamorphosed than the older rocks of the crest!

It is unfortunate that thorough studies of the Paria-Araya metamorphics have not been made since their age, whether Upper Cretaceous, Lower Cretaceous, early Mesozoic, or older, must remain a moot question until observational facts can be substituted for speculation.

## Cretaceous Strata (Northern Mountain Front)

### *Sucre Group*

### (Fig. 5; Pl. 6)

*General discussion.*—Cretaceous sediments constitute the northern and principal portion of the Serranía del Interior of Eastern Venezuela. Three formations are usually assigned to the Lower Cretaceous. The oldest (Barranquín) consists of an alternation of continental sediments, coastal swamp deposits, and shallow-water marine sediments, showing more persistently marine conditions toward the top of the formation. The next younger formation (El Cantil) consists dominantly of reef limestones with some dark shales. The uppermost unit (Chimana) is an alternation of shallow-water marine sandstones, limestones, shales, and glauconitic rocks somewhat similar to the upper part of the Barranguín formation. These Lower Cretaceous formations are land or shallow marine facies of the same general cycle of deposition and are both laterally and vertically gradational with each other. They are all three here included in a single group for which the name *Sucre group* is suggested because of their excellent development in the State of Sucre. The difficulties in sharply differentiating the individual formations of the group have been stressed recently by Gonzales de Juana (1947, p. 695).

The Sucre group is overlain with apparent conformity by the Upper Cretaceous Guayuta group of carbonaceous-bituminous shaley limestones and limey shales representing a deeper-water and laterally more persistent environment. Although no unconformity is discernible a stratigraphically condensed interval lacking in diagnostic fossils probably intervenes between the Lower and Upper Cretaceous groups. This interval probably corresponds approximately to Upper Albian and part of Cenomanian time. The Upper Cretaceous Guayuta group is overlain conformably by a shallower-water more clastic series of sediments of Late Cretaceous to Tertiary age included under the term of Santa Anita group.

The Cretaceous formations are known in

typical development along the northern border of the Eastern Venezuela Basin as far west as the coastal islands of Guanta. They are largely covered by the sea opposite the Barcelona gap (Barcelona to Unare River) and are but poorly known farther west in the Serranía del Interior of Central Venezuela due to structural complications, overlap, and possible transition to metamorphic rocks.

*Barranquín formation* (Pl. 6, fig. 1).—The type locality was designated by Liddle (1928, p. 109–111) as the vicinity of the town of Barranquín, State of Sucre, on the Aragua de Maturín-Cumaná trail from Dos Ríos on the south to within 2 kilometers of San Fernando on the north. The formation includes a variety of lithologic types but consists dominantly of coarse, cross-bedded sandstones and variously colored shales. Thin yellowish-brown limestones are common in the upper part.

This formation makes up most of the large area of Cretaceous rocks in northeastern Venezuela which composes the Serranía del Interior of this region. It is particularly well developed in the northern part of this area where it forms the broad belt of mountains and rugged hills extending from the lowlands bordering the Gulf of Paria westward to the islands off the Guanta coast. It is also locally well exposed near the southern border of the Serranía in its eastern portion between Aragua de Maturín and Caripito where it is brought up on such major structures as Punceres and Cerro García. Liddle estimates an exposed thickness of 3500–4000 feet, and a thickness of 4900 feet has been measured near Aragua de Maturín. However, a normal sedimentary contact with older formations has not been found, and the total thickness is therefore not known since it is not certain that the basal part of the formation is exposed anywhere in this region. The northern edge of the Barranquín formation outcrop area is formed by the Gulf of Cariaco and the so-called Cariaco depression which extends eastward through northeastern Venezuela to the Gulf of Paria. These separate the unmetamorphosed Barranquín formation on the south from the metamorphics of Araya and Paria on the north and probably reflect a major zone of faulting. However, the actual contact is hidden by the waters of the gulf or the alluvial fill of the lowlands except in a very limited area where a connecting link of higher ground exists between the Serranía del Interior and the Serranía de la Costa. Here, in the vicinity of the town of Río Casanay, unmetamorphosed fossiliferous Barranquín limestones, shales, and sandstones are in abrupt contact with the metamorphic rocks which consist of dense blue marbleized limestones, schists, and phyllites. A prominent breccia (and a warm spring) occurs at the contact and supports the idea that this is a major fault contact.

The Barranquín formation in the southeastern part of the Serranía del Interior near Aragua de Maturín and Punceres is composed of coarse and gritty cross-bedded light-colored quartzitic sandstones, variously colored (red, purple, brown, yellow, gray, and bluish) sandy shales, black carbonaceous shales, and a few beds of hard gray platy limestone. Individual sandstones as much as 200 feet thick occur in the middle portion of the exposed sections. The formation as seen in the central part of the Serranía near the peak of Turumiquire and near Cumanacoa consists predominantly of massive to thin-bedded, cross-bedded, coarse, frequently gritty, quartz sandstones. The interbedded shales are usually sandy and micaceous. They are probably actually gray or dark gray but generally are weathered to reds, browns, and purples. Bluish-gray reef limestones are common in the upper part of the formation and reach thicknesses of 20–30 feet. Many of these are pebbly. They are frequently fossiliferous. Good exposures occur all along the coastal road from Guanta to Cariaco. Here also reddish-weathering quartzitic sandstones and grits predominate, but carbonaceous plant-bearing shales and yellowish-weathering blue-gray fossiliferous limestones are common.

The upper part of the formation is beautifully exposed on many of the islands off the coast near Guanta. Characteristic lithologic types noted here (Hedberg and Pyre, 1944) are:

1. Light-colored, reddish-weathering, coarse-grained quartzitic sandstones. Highly cross-bedded. Commonly conglomeratic with quartz pebbles. Locally whitish due to decomposed feldspar.
2. Shaly, highly micaceous, fine- to medium-grained sandstones.
3. Dark-gray carbonaceous shales with well-preserved plant remains and gray splintery

micaceous shales weathering to lavender, purplish, or reddish.

4. Gray, olive-brown weathering, fossiliferous, sandy limestones with fossils commonly preserved as iron-oxide replacement. Rudistid reef limestones.

The outcrop areas of the Barranquín formation are very characteristically reddish due to weathering. This is readily discernible from the air over much of the area.

The writer knows of no typical Barranquín formation west of Barcelona in the Barcelona-Unare gap or along the southern flank of the Serranía del Interior of Central Venezuela. Probably, however, rocks of this age are included in the metamorphic series of these mountains.

As discussed previously, the base of the formation has not been seen. The formation grades upward conformably into the El Cantil or El Cantil-Chimana formations, and the boundary must frequently be drawn very arbitrarily to separate a section in which sandstones and plant-bearing shales are dominant (Barranquín) from one in which limestones and marine shales are dominant (El Cantil). Individual limestones in the upper Barranquín are frequently indistinguishable from limestones in the El Cantil, and likewise individual sandstones in the lower El Cantil are indistinguishable from those in the Barranquín. Because of lateral variation in lithologic facies the contact probably varies considerably in time-stratigraphic position from place to place.

Fossil occurrences in the formation have been summarized previously (Hedberg, 1937a, p. 1983; Hedberg and Pyre, 1944, p. 7). A review of the paleontological data shows that the marine fossiliferous upper part of the formation has strongest affinities to the Aptian and Barremian-Aptian (Urgonian) stages of Europe. It is not improbable that the lower and less marine part extends down into the Neocomian stage.

The Barranquín formation represents the beginning of the Cretaceous transgression in Eastern Venezuela. In its lower part it is probably largely nonmarine, but passes upward into an alternation of shallow-water marine beds with nonmarine coastal swamp deposits, with conditions becoming more persistently marine toward the top of the formation.

Approximate time and facies equivalents of the Barranquín formation include the Segovia series of Dalton in the Carora-Barquisimeto region, the Tomón formation of the Venezuelan Andes, the Uribante formation of Táchira and Colombia, and the Río Negro formation of the Perijá mountains. Some of the formation may be equivalent to the lower part of the Temblador formation of the southern edge of the Eastern Venezuela Basin.

*El Cantil formation* (Pl. 6).—The formation was named by Liddle (1928, p. 124–135) from the locality of El Cantil near the village of Mesa de Punceres in the State of Monagas. The type section is exposed on Río Punceres on the south flank of the Punceres anticline. Liddle indicates that the formation here consists of about 800 feet of limestone and shale with the former predominating by three to one. However, the writer cannot reconcile the exposures found on a recent traverse of the Río Punceres with Liddle's description of this type section of the formation. The section appears to be badly faulted so as to make it difficult to be sure of more than a few hundred feet of continuous sequence, and there appear to be coarse, glauconitic, cross-bedded Barranquín-like sandstones, as well as dark shales within the typical El Cantil limestone section. This is a very poor type section, and it is difficult to define the formation in terms of the exposures here. Through usage, however, the formation name has come to be applied to a dominantly massive gray limestone development between the Barranquín sandstones and the overlying Guayuta group. Where marine sandstones and/or shales largely replace the typical gray and blue-gray reef limestones the unit is perhaps better called Chimana formation. (*See* discussion of Chimana formation.)

Outcrops of the El Cantil formation extend throughout the whole east-west extent of the Serranía del Interior of Eastern Venezuela and form some of the highest peaks of the range. Massive limestones of this formation form the walls of the famous Guácharo cave near Caripe. At Cerro García north of Aragua de Maturín a thickness of 1500 feet was measured. The lower part of the formation here is an alternation of thin limestones and shales with occasional sandstones. The upper part includes thicker and more massive limestones, and the formation as a whole is more than 50 per cent limestone. Glauconitic beds are found near the top. Sand-

stones are more common throughout the formation than is usually recognized.

Liddle (1928, p. 124–135) has described numerous occurrences of the formation throughout the Serranía del Interior. The outcrop belt in general lies south of the Barranquín formation outcrop, but fold and fault structure have resulted in numerous irregularities in areal distribution.

In the northeastern Anzoátegui the formation is particularly well exposed near Guanta on the road to Cumaná and on the islands off the coast. At the eastern end of the island of Chimana Grande the formation is about 1200 feet thick. Here the upper 500 feet consists almost entirely of very resistant massive dense bluish-gray "reef" limestone which weathers to characteristic cannelated and pinnacled forms (Pl. 6, fig. 3). The lower 700 feet consists of dark fissile shale with some thin interbedded yellowish fossiliferous limestones. This lower shaly facies, however, grades westward across the island into massive limestones, and throughout the islands the formation is characterized by rapid lateral changes. The shales of the formation are usually poor in Foraminifera as contrasted with Upper Cretaceous shales.

Farther west at Morro de Unare rudistid limestones have been reported which may belong to this formation. Other remnants of rudistid limestone which also may belong to it are associated with faults along the southern edge of the Serranía del Interior of Central Venezuela at Batatal, San Francisco de Macaira, and elsewhere. However, at most the formation is very poorly represented in the nonmetamorphosed sediments at the edge of this part of the basin. Marmorized limestones farther north within the metamorphosed portion of the Serranía del Interior of Central Venezuela may represent the El Cantil formation.

The contact between the El Cantil formation and the underlying Barranquín formation is placed rather arbitrarily where the dominantly sandstone and shale lithology of the Barranquín gives way to a section in which limestones or interbedded limestones and marine shales predominate. The contact represents a change toward more persistently marine conditions. As a result of rapid lateral variation in lithology, the stratigraphic position of the contact probably changes considerably from place to place. The formation is conformable both with the Barranquín formation below and with the overlying strata of the Chimana formation (or Querecual formation where Chimana formation is absent due to lateral intergradation with El Cantil).

The formation is at least locally very fossiliferous. Paleontologic evidence on its age has been reviewed briefly (Hedberg and Pyre, 1944, p. 8). These data together with the age determinations for the overlying and underlying formations indicate that the El Cantil formation is largely Aptian and lower Albian.

The El Cantil formation is entirely marine. It was probably deposited in relatively shallow water, and the "reef" or bioherm character of the massive limestones of the upper part is evident from the rapid lateral variation to clastic sediments, the massive structureless character of the rock, and the local abundance of characteristic bioherm-forming fossils such as corals and *Orbitolina*.

The El Cantil formation is similar in facies and stratigraphic position to the Cogollo formation of the Venezuelan Andes and Maracaibo Lake Basin and to the Capacho formation of Táchira. It is probably approximately the time equivalent of these formations also. A number of scattered occurrences of rudistid limestone in Trinidad are also probably approximately equivalent in age to the El Cantil formation.

*Chimana formation* (Pl. 6).—This formation was named from the island of Chimana Grande, 7 kilometers north of Puerto La Cruz (Hedberg and Pyre, 1944, p. 8–10). The type section is almost perfectly exposed at the eastern end of the island on the south slope of the high east-west El Cantil limestone ridge, and excellent exposures continue throughout the length of the island. The name Chimana formation replaces the provisional term "Bergantín beds" used for this formation in a previous publication (Hedberg, 1937a, p. 1985).

The type section consists largely of fissile micaceous shale weathering to lavender, maroon, yellowish-gray, orange, and olive-brown tints. There are some discontinuous olive-brown fossiliferous limestones in the lower part of the formation and some thin seams of reddish ironstone interbedded in the shales. Brownish-weathering glauconites and glauconitic mica-

ceous sandstones are common in the upper part. There is 400 feet of Chimana formation in the type section, but the formation ranges from 200 to 1000 feet thick due to lateral gradation to El Cantil facies.

The formation is well exposed on other islands in the group near Guanta and Puerto La Cruz, but closely adjacent sections show great differences due to rapid lateral lithologic variation within this interval between the massive reef limestone of the El Cantil formation and the dark limestones and shales of the overlying Querecual formation.

On the mainland near Guanta, the formation is in general similar to that of the islands. Farther east and along the southern edge of the Serranía del Interior the formation has been tentatively identified in the crestal areas of the Bergantín and Mundo Nuevo anticlines, respectively, some 40 and 60 kilometers east-southeast of Barcelona. Here there is about 1000 feet of shales, glauconitic micaceous sandstones, and thin *Orbitolina* limestones underlying the Querecual formation which may be considered as belonging to the Chimana rather than the El Cantil facies. In parts of this area the whole El Cantil formation may be missing due to lateral gradation to Chimana facies. Farther east in the Serranía del Interior the Chimana formation, if present, has not been separated from the El Cantil formation. The formation has not been definitely identified west of Barcelona, but its time equivalents occur as fragments in the tectonically disturbed frontal part of the Serranía del Interior of Central Venezuela.

The formation in typical development is not very resistant, and where the underlying El Cantil limestone is present the Chimana formation is usually found on the slopes of El Cantil hills or as valleys between the El Cantil and Querecual formations. On the Bergantín and Mundo Nuevo structures, it forms anticlinal valleys along the crests of these folds. The outcrop area usually weathers to a deep reddish due to the abundance of glauconite.

Where the El Cantil limestone is well developed, the lower boundary of the Chimana formation may readily be placed at the point where the characteristic thick massive dense gray reef limestones of the El Cantil give way to the shales, glauconitic micaceous sandstones, and thin olive-brown limestones of the Chimana formation. However, since the El Cantil reefs frequently grade laterally into sediments of the Chimana facies and vice versa, it is often difficult to decide on an exact contact. Thus on La Borracha island a massive "reef" limestone more than 100 feet thick develops rapidly in the lower part of the Chimana formation. This limestone is in itself essentially indistinguishable from the main El Cantil limestone, but is separated from it by some 200 feet of typical Chimana shales and sandstones.

The contact of the Chimana formation with the overlying Querecual formation is one of the sharpest and most readily determined formational contacts in northeastern Venezuela. The change from the shallow-water sandstones, limestones, glauconite beds, and varicolored weathering shales of the Chimana formation to the uniform dark-gray or black limey shales and shaly limestones of the Querecual formation is striking and takes place in a transition shale zone only about 50 feet thick (Pl. 7, fig. 4). The abruptness of the contact and the age difference between the two formations as indicated by fossils have suggested an unconformity, but these features appear to be due to a rapid change in environment when deposition was slow rather than to any actual emergence.

Fossil occurrences in the Chimana formation have been discussed previously (Hedberg, 1937a, p. 1986–1988; Hedberg and Pyre, 1944, p. 9–10). On the basis of available fossil evidence it appears that the Chimana formation, like its lateral equivalent the El Cantil formation, is of Aptian-Albian age. In its upper part it is probably Albian and may even reach to the Cenomanian although there is no paleontologic evidence of upper Albian. At the same time it is interesting to note that the pelecypods, gastropods, and corals from the lower part of the Chimana formation very closely resemble those of the upper part of the Barranquín formation of Urgonian age, paralleling the close similarity in lithologic facies of the two formations. The possibly misleading influence of mere similarity of depositional facies on conclusions regarding correlation and age determination in these shallow-water sediments is worthy of serious thought.

In an effort to determine as closely as possible the upper time limit of the Chimana beds, a de-

tailed collection of fossils was made at a horizon only 50–100 feet below the contact with the Querecual formation and sent to R. W. Imlay for determination. On the basis of *Oxytropidoceras*, *Lyelliceras*, and other ammonites he believes this horizon is definitely no younger than middle Albian.

The Chimana formation throughout is a shallow-water marine deposit, as indicated by the coarse micaceous sandstones, shallow-water mollusks, scarcity of small Foraminifera, abundance of glauconite, and rapid lateral gradation to reef limestones. It is to be correlated in age with the upper part of the El Cantil formation and its equivalents.

### *Guayuta Group*

(Fig. 5; Pl. 7)

*General discussion.*—The term Guayuta (Liddle, 1928, p. 154) is now used as a group term to include that part of the thick series of Upper Cretaceous sediments in Eastern Venezuela which is dominantly characterized by thin-bedded black carbonaceous-bituminous shaly or cherty limestones and calcareous shales (Hedberg, 1937a, p. 1989–1992). Finely laminated limestones and discoidal limestone concretions are also characteristic. This is a remarkably uniform and widespread stratigraphic unit. It is probably 3000–4000 feet thick and extends more or less continuously throughout the southern part of the Cretaceous area of northeastern Venezuela (Fig. 4), from the Guanoco asphalt lake on the east to the vicinity of Puerto La Cruz on the west. It is not exposed in the vicinity of the Barcelona-Unagre gap, but thin shreds of the unit are known along the faulted front of the Serranía del Interior of Central Venezuela in northwestern Anzoátegui and northeastern Guárico.

The Guayuta group was named from the Río Guayuta between Aragua de Maturín and Punceres in the State of Monagas (Liddle, 1928, p. 134, 161). The upper portion of the group is beautifully exposed in the river gorge here and consists of thin-bedded black carbonaceous-bituminous shaly and cherty limestone, and calcareous shale. The cherty character particularly predominates in the upper part. Thin sandstones appear near the top, and the unit grades upward conformably into the sandstones and sandy shales of the Santa Anita group. The writer cannot find any indication of the angular unconformity described by Liddle. The contact of the Guayuta group with the subjacent El Cantil formation is probably a fault contact.

On Río Querecual near Bergantín the Guayuta group was divided into the Querecual formation below and the San Antonio formation above (Hedberg, 1937a, p. 1929–1994; Hedberg and Pyre, 1944, p. 10–12). The basis for separation is the presence of sandstones in the upper part of the group (San Antonio) and its more cherty character. The distinction can be made quite readily throughout the whole of the Serranía del Interior all the way to its eastern end where the so-called Guanoco shale corresponds to the San Antonio formation. The group term is useful, however, when isolated exposures or transitional characters make difficult the differentiation into individual formations.

*Querecual formation* (Pl. 7).—The formation is very well exposed in its type section on Río Querecual near Bergantín in northeastern Anzoátegui (Hedberg, 1937a, p. 1989–1992). It consists almost entirely of hard black thin-bedded and laminated carbonaceous-bituminous shaly limestone and calcareous shale. Discoidal, spheroidal, and ellipsoidal limestone concretions ranging from a few inches to several feet in diameter are particularly characteristic (Pl. 7, fig. 1). The finely laminated character of the Querecual limestone is very persistent and is apparently due to a fine alternation in the relative proportions of calcium carbonate and carbonaceous matter. There are frequently 10–20 laminae to the inch. Thin-section study of the limestones shows that many of them are made up in large part of the tests of planktonic Foraminifera (*Globigerina*, *Gümbelina*, etc.). The rock commonly gives a strong petroliferous odor when freshly broken. Some black chert intervals occur in the upper part of the formation. In open areas the formation weathers to a whitish gray contrasting with the reddish weathering of the older Cretaceous formations. On Río Querecual the formation is about 2400 feet thick, but, in general, measurements of this formation are very difficult owing to its incompetence and susceptibility to structural deformation. Although individual beds of the

typical Querecual formation are less resistant than the sandstones, limestones, and cherts of adjacent formations, the great thickness of homogeneous though only moderately durable material of which the formation is composed has frequently resulted in strong topographic features throughout the Serranía del Interior.

As previously discussed the contact with the underlying Chimana or El Cantil formations is remarkably distinct although the two formations appear to be conformable. The division line between the Querecual formation and the overlying San Antonio member of the Guayuta group is placed at the base of the first more or less continuous sandstone bed in the group, and the distinction between the two formations is made largely on the presence of sandstones and more abundant chert beds in the latter. Where exposures are poor this contact may be difficult to draw.

Fossil occurrences have been summarized previously (Hedberg and Pyre, 1944, p. 11) and point to a Turonian age for this formation. Recent work by Imlay on material from the lower part of the formation on the islands off the Guanta coast has shown nothing older than Turonian. Since a horizon near the top of the underlying Chimana formation is considered by him as not younger than middle Albian, the upper Albian and Cenomanian if present must be represented by a relatively condensed section in the thin transition interval between the two formations.

*San Antonio formation* (Pl. 7).—This formation constitutes the upper part of the Guayuta group. The type section is on the Río Querecual south of Bergantín in northeastern Anzoátegui, and the formation is named from Cerro San Antonio near this town (Hedberg, 1937a, p. 1992). The formation consists largely of dark-gray carbonaceous-bituminous shaly and cherty limestone and calcareous shale of the type characteristic of the underlying Querecual formation. However, it is generally more cherty and contains a number of beds of massive light-gray well-sorted fine- to medium-grained calcareous sandstone ranging from 1 to 35 feet thick. These sandstones are very discontinuous laterally. Sandstone dikes ranging from a few inches to 5 feet thick are also particularly characteristic (Pl. 7, fig. 3). The type section of the formation is about 1300 feet thick.

The San Antonio formation with its characteristic chert and sandstone beds is well exposed on the Aragua de Maturín-Cumanacoa road just north of the principal divide and is a persistent unit from here to the western end of the Serranía. The chert development is quite variable laterally. In the Barcelona area at Morro de Barcelona and on La Borracha Island the entire formation is chert or siliceous limestone.

The so-called Guanoco formation outcropping near the asphalt "lake" of Guanoco in the State of Sucre at the eastern end of the Serranía del Interior consists largely of brown and black calcareous and frequently highly siliceous shale weathering to a yellowish or whitish color and punky character. There are some sandstones and dark limestones. The unit is probably approximately equivalent to the San Antonio formation.

The resistant chert and sandstone beds of the San Antonio formation are usually the dominating element in the development of Guayuta group topography. This topography is generally quite characteristic with steep precipitous slopes and strong V-shaped development of both valleys and ridges.

In the Río Querecual section and farther east the upper limit of the San Antonio formation (and Guayuta group) is readily identified by the thick gorge-forming San Juan sandstone which constitutes the basal member of the overlying Santa Anita group. In the Barcelona area where the San Juan sandstone is missing or represented only by thin sandstones separated by shales, the contact is less clear and must be placed rather arbitrarily where the characteristic Guayuta group lithology of carbonaceous-bituminous limestones, calcareous shales, and cherts gives way to sandstones and noncalcareous or only slightly calcareous foraminiferal shales.

Fossil occurrences in the San Antonio formation have been discussed previously (Hedberg, 1937a, p. 1993–1994; Hedberg and Pyre, 1944, p. 12). While the San Antonio, like the Querecual, is characterized by planktonic Foraminifera, benthonic forms have been found in considerable abundance in the shaly portions of the uppermost part of the formation and include a number of species common to the lower part of the Colon shale of western Venezuela and

Colombia. The San Antonio formation is probbably in large part early Senonian.

The Guayuta group in general is the eastern Venezuelan representative of that widespread and distinctive facies of Upper Cretaceous sedimentation which formed the La Luna formation of western Venezuela, and the Villeta and Palmira formations of Colombia, and which is known from Trinidad to Ecuador and perhaps farther south. The more siliceous facies of these sediments are known locally as the San Antonio (and Guanoco) formation (Eastern Venezuela), Táchira chert (Táchira and Santander del Norte), Guadalupe formation (Colombia), and Chert Hill beds (Trinidad).

The Querecual member of the Guayuta group appears to represent an abrupt change from the aerated waters of the shallow marine environment of Lower Cretaceous time to a more quiet and probably deeper sea where, owing to restricted circulation, benthonic life was largely inhibited and plankton together with the remains of free-swimming life (fish, ammonites) accumulated undisturbed. Much of the Querecual sediment at the time of deposition must have looked like a veritable *Globigerina* mud or ooze. Clastic sediment was limited to fine clay or mud particles and to carbonaceous matter. During San Antonio time there was a gradual but persistent change from the specialized environment represented by the Querecual formation. A greater amount of clastic material was brought into the seas of this area resulting in local sandstone developments. Better circulation and aeration of the marine waters allowed the development of benthonic foraminiferal faunas from time to time partially supplementing the planktonic Foraminifera of the Querecual. Finally, extensive deposits of chert and siliceous shale indicate an increased silica content of the in-flowing waters as well as conditions favorable for the precipitation of silica. The belief that the chert of the San Antonio formation is of primary origin is supported by its geographic extent, the fact that it is confined to a fairly definite stratigraphic interval, and the preservation within it of abundant calcareous Foraminifera.

The thin-bedded relatively incompetent sediments of the San Antonio and Querecual formations (Guayuta group), lying as they do between the more massive competent units of the El Cantil and Santa Anita formations, have frequently been highly crumpled and distorted. The resulting structural discrepancy between the Guayuta beds and the massive overlying San Juan sandstone of the Santa Anita formation has sometimes been erroneously interpreted as evidence of angular unconformity.

## Cretaceous Strata (Southern Facies)

### *Temblador Formation*

(Fig. 5)

*Description.*—On the southern limb of the Eastern Venezuela Basin in Southern Monagas, Anzoátegui, and Guárico, wells drilled for oil have encountered Cretaceous sediments underlying the Oligocene-Miocene Oficina formation. These sediments were given a formation name from the Temblador oil field in southeastern Monagas where they were first encountered, although the best reference section is the completely cored section of the Tigre No. 1 well in southeastern Anzoátegui. The formation is defined and discussed in detail in a recent paper on the oil fields of the Greater Oficina area, Central Anzoátegui, Venezuela (Hedberg *et al.*, 1947).

The Temblador formation is not known in outcrop, and its description is based on well sections. It has been divided into an upper and lower member:

*Upper glauconitic member.*—This member includes 300 to 600 feet or more of gray, greenish-gray, and green glauconitic fine-grained sandstones and siltstones, gray friable coarse and gritty sandstones, dark carbonaceous phosphatic shales, lignitic shales, whitish powdery sandstones and siltstones, and rare beds of hard gray massive glauconitic and usually fossiliferous dolomites, dolomitic limestone, and dolomitic sandstone. There are a few intervals of whitish and mottled claystones.

*Lower mottled member.*—The lower member ranges from 300 to 800 feet thick and is made up principally of coarse grits, sandstones, siltstones, and claystones, all usually mottled in dull shades of light gray, greenish gray, yellow,

brown, red, and purple. Whitish powdery grits, sandstones, and siltstones are common here as well as in the upper member and there are some intervals of bluish-gray claystone with plant remains. The division between the two members is drawn principally on the absence of glauconite and predominance of mottled beds in the lower member.

The Temblador formation rests with profound unconformity on an apparently peneplained surface of igneous and metamorphic basement rocks and Jurassic-Triassic (?) Carrizal and Hato Viejo formations. Wherever known it is overlain by the Oligocene-Miocene Oficina formation. In spite of the time gap at this upper contact there is surprisingly little evidence of unconformity; and especially where the two formations are dominantly sandy careful laboratory work is required to determine the contact accurately.

The Cretaceous age of the formation was definitely established by the finding of species of *Exogyra*, *Plicatula*, *Astarte*, *Neithea*, and *Turritella* in the upper member in the Tigre No. 1 well section. Additional Cretaceous fossils have been found in other wells. Dark shales in the formation characteristically yield linguloid brachiopods and fish remains. Fossils from the Tigre No. 1 section were submitted to R. W. Imlay who believes they are not older than Cenomanian.

The lower Temblador is evidently a nonmarine deposit derived directly from erosion of the Guayana shield igneous and metamorphic complex. The upper Temblador deposits, on the other hand, represent an alternation of brackish, lagunal, and coal-swamp conditions with a true shallow-water marine environment.

The Temblador formation is most certainly equivalent to some part of the outcropping Cretaceous section of the Serranía del Interior to the north, but its exact time relation to those sediments is difficult to determine accurately because of the distance, scarcity of fossils, and difference in facies. Although lithologically most similar to the Lower Cretaceous section, the Temblador is more probably the near-shore and continental facies of the marine Upper Cretaceous (Guayuta and Santa Anita) farther north.

## Cretaceous—Tertiary Strata

### *Santa Anita Group (Northeastern Mountain Front)*

(Fig. 5; Pl. 8)

*Description.*—The term Santa Anita was originally used as a formation name to include the sediments between the top of the Cretaceous Guayuta group and the base of the Upper Eocene Merecure formation on Río Querecual (Hedberg, 1937a, p. 1994). The formation was later divided into three members in northeastern Anzoátegui (Hedberg and Pyre, 1944, p. 12–15). Marked lithologic variations within this interval of sediments throughout the Serranía del Interior and elsewhere have given rise to recognition of other stratigraphic units, and it is here proposed to raise the term to the rank of a group which will include as formations those members defined in northeastern Anzoátegui as well as other units which may later be formally established. The Santa Anita group includes a portion of what Liddle (1928, p. 216–225) included in his "Aragua sandstone."

The Santa Anita group consists of several thousand feet of quartzitic sandstones, dolomitic and calcareous sandstones and siltstones, gray shales, and glauconitic rocks forming a more or less continuous series of outcrops along the southern porder of the Serranía del Interior of Eastern Venezuela from Puerto La Cruz on the west to Río Guayuta near Aragua de Maturín on the east. Coarse clastics of conglomerate grade are rare. Eastward from Río Guayuta younger Tertiary sediments completely overlap the Santa Anita group against older Cretaceous strata brought up on folded structures at the southeastern edge of the Serranía. Liddle (1928) reports a small thickness of "Aragua sandstone" locally outcropping near Caripito and near Guanoco, but we lack sufficient evidence to show this as belonging to the Santa Anita group.

The outcrop area of the group is characterized by rugged sandstone ridges with intervening shale valleys, and the basal sandstone holds up many of the most elevated points at the southern edge of the Serranía. Since the group includes both Cretaceous and Eocene strata, the

Cretaceous-Eocene boundary on Bucher's Geologic-tectonic Map of Venezuela (Bucher, 1950) falls within its out-crop belt.

Sediments of the Santa Anita group are missing in the Barcelona gap west of Barcelona, but from the vicinity of Río Unare westward sediments of this group are present all along the northern edge of the basin.

*San Juan sandstone.*—The type section is exposed on Río Querecual and consists of 320 feet of very hard gray to light-gray, fine- to medium-grained, well-sorted, sparingly glauconitic, and locally calcareous sandstone in beds ordinarily 1–3 feet thick. In the type section this formation constitutes the base of the Santa Anita group. Westward from Río Querecual the formation grades within a short distance into shale with only minor beds of sandstone and is included in the Vidoño shale formation. Toward the east, however, the San Juan sandstone thickens at the expense of the Vidoño shale to more than 1000 feet of fine- to coarse-grained sandstone with only minor breaks of shale. Here it constitutes one of the most resistant units of the southern edge of the Serranía del Interior forming high peaks and ridges and difficultly passable gorges on the main rivers.

The San Juan formation is known as far east as Aragua de Maturín where a large part of the thick "Aragua sandstone" of Cerro Aragua belongs to this formation. The cherty beds and thin sandstones of the upper part of the Guayuta group in Río Guayuta grade upward transitionally on the flanks of Cerro Aragua through a section of alternating medium-bedded and thin-bedded sandstones and sandy shales to the thick series of massive fine- to coarse-grained sandstones which form the upper part of the hill. The unit here exceeds 2000 feet.

The San Juan formation is definitely Upper Cretaceous as proved by the foraminiferal fauna of the immediately overlying Vidoño shale. The contact with the underlying San Antonio formation of the Guayuta group is frequently abrupt, but there is no reliable indication of unconformity. On the contrary, sections like that of Cerro Aragua show a complete transition from one to the other. Apparent angular discrepancy, which has sometimes been reported, is actually the result of the relative incompetence of the underlying Guayuta beds as compared with the resistant San Juan sandstone.

*Vidoño shale.*—This formation is named from the village of Vidoño, 6 kilometers east of Barcelona, which is situated in a valley formed by this shale. A good reference section is exposed on Río Querecual and its tributary Quebrada San Juan (Hedberg and Pyre, 1944, p. 12). The formation consists almost entirely of dark-gray foraminiferal shale with a few minor beds of sandstone and hard calcareous siltstone. All the lithologic types are commonly highly glauconitic. The unit is about 900 feet thick on Río Querecual but thickens westward. In the vicinity of Barcelona and on La Borracha Island it is in contact with the Guayuta group due to pinching out of the San Juan sandstone. Eastward from Río Querecual the formation thins until it may be considered as grading into the San Juan and Caratas formations. The formation is generally expressed topographically as a valley between the San Juan or Guayuta formations below and the Caratas formation above.

The Vidoño shale carries a good foraminiferal fauna (Hedberg and Pyre, 1944, p. 13–14; Cushman, 1947). In its lower part it is approximately of upper Senonian age. Its upper part carries a fauna of Cretaceous aspect but lacking in species definitely known to be restricted to the Cretaceous. There is considerable similarity between this fauna and that of the Lizard Springs beds of Trinidad, which is considered of Maestrichtian-Danian age by Cushman and Renz (1946). The formation is conformable with both underlying and overlying formations and is defined principally as a dominantly shale unit between more resistant beds.

*Caratas formation.*—The type section of this formation is on Río Querecual (Hedberg and Pyre, 1944, p. 13). Here it includes about 1100 feet characterized principally by distinctive black or blue-gray very hard calcareous and dolomitic siltstones and impure silty and commonly dolomitic limestone (Pl. 8). It also includes a 130-foot interval in the midst of the formation characterized largely by very hard fine- to medium-grained sandstone. There are several large shale intervals. Glauconite is common throughout. The formation is in contact

above with sediments of the Upper Eocene Merecure group.

The contact of the Caratas formation with the Merecure group is in general placed at the top of the distinctive calcareous or dolomitic siltstones of the former and at the base of the prominent Los Jabillos sandstone formation or the Tinajitas orbitoidal limestone formation of the latter. In the Barcelona area the Caratas formation is only a few hundred feet thick, probably as a result of erosional unconformity and the contact is sharp and easily placed. (Pl. 8, fig. 3.) This is also the case as far east as Río Capiricual.

On Río Querecual there is about 100 feet of shaly to gritty calcareous concretionary sediments which apparently forms a transition zone between Caratas formation and the Merecure group. Here there appears to have been no major unconformity developed, but there was shallow-water deposition with local discontinuities of sedimentation or even local emergences above the base level of deposition. Farther east along the southern edge of the Serranía del Interior the Caratas formation thickens and becomes increasingly sandy with even some gritty and pebbly beds in its upper part. On Río Aragua, Río Orégano, and other streams, there is a transition zone between the two units in which dense black calcareous siltstone typical of the Caratas formation alternates with pebbly quartzitic sandstone like that of the Los Jabillos formation of the Merecure group. It is thus probable that in this local area from Río Querecual to Río Guarapiche in the foothills of the Serranía del Interior there was no break in deposition between the Santa Anita and Merecure groups such as exists to the west and must be present farther south. The Caratas formation has not been identified as far east as Aragua de Maturín.

Shaly portions of the Caratas formation carry a meager fauna of small Foraminifera of which most of the species are also present in the underlying Vidoño shale. Calcareous glauconitic grits and siltstones and thin limestones near the top of the formation carry nummulites and orbitoids (*Discocyclina*, *Lepidocyclina*, *Miscellanea*). A limestone near Cerro Corazón north of Urica, believed to belong in the Caratas formation, carries abundant specimens of a planicostate *Venericardia* suggesting Paleocene age (Pl. 8, fig. 2). (*See* discussion of fossil fauna from this formation, Hedberg and Pyre, 1944, p. 14.) The Caratas formation of the Barcelona area is probably Paleocene, but farther east where it is transitional upward into the Merecure group it may include sediments of lower and middle Eocene age which have been removed by erosion in the Barcelona area.

Sediments of the Santa Anita group were deposited largely in a shallow variable sea marking the close of the cycle of deposition which had started in early Cretaceous time. In the eastern part of the area where the San Juan formation was deposited much clastic sediment was being brought in, probably from the south. Farther west deep-water conditions persisted longer as indicated by the Vidoño shale deposits. In Caratas time shallow-water marine deposition prevailed throughout the length of the Serranía with local areas of limestone reef formation. Peculiar conditions of sedimentation are represented by the dark dolomitic and calcareous siltstones characteristic of this formation. Slow deposition on a static sea floor is suggested by the large amount of Eocene time believed to be represented by these deposits.

The Santa Anita group is largely of uppermost Cretaceous age (Maestrichtian-Danian). However, the orbitoids and nummulites of the upper part of the Caratas formation indicate that it extended well into Eocene time. The apparent transition in the eastern part of the area into Upper Eocene Merecure sediments would require that it represent not only Paleocene but also both lower and middle Eocene time.

The Vidoño shale is approximately equivalent in age to the Upper Colón and Mito Juan formations of Western Venezuela and the Lizard Springs formation of Trinidad. The *Venericardia* limestone of the Caratas formation appears to represent the Soldado horizon of Trinidad and the Guascare horizon of Western Venezuela. The upper part of the Caratas formation known from the eastern part of the Serranía del Interior should be approximately equivalent to the Navet formation of Trinidad. The Caratas formation in general appears to be equivalent to the Misoa-Trujillo formation of Western Venezuela.

### *Santa Anita Group (Northwestern Mountain Front)*

(Fig. 5)

Between Barcelona and Píritu no sediments of the Santa Anita group are exposed. However, west of Píritu near Río Unare several hundred feet of dark-gray to bluish-gray foraminiferal shale with some calcareous glauconitic sandstone is known which, on the basis of the Foraminifera, may be considered Vidoño shale. The Caratas formation is apparently missing.

West of Río Unare along the highly faulted front of the Serranía del Interior of Central Venezuela isolated outcrops of shales with a Vidoño formation fauna are known as far east as San Francisco de Macaira.

At a number of points along the southern part of the Serranía del Interior from the Guárico-Anzoátegui boundary westward are isolated reef limestone hills or *morros* (Batatal, San Francisco de Macaira, San Sebastián, San Juan de los Morros, etc.). These limestones are sometimes characterized by a distinctive pseudo-oölitic structure. Faulting or lack of exposures makes their relation to adjacent rocks difficult to determine. Caudri (1944) has determined the fauna in association with some of these *morros* near San Juan and near San Sebastián as Paleocene and approximately equivalent to the Soldado formation of Soldado rock and Trinidad. She has called attention, however, to certain rather anomalous data which might have suggested Cretaceous age. These limestones probably belong to the Santa Anita group and appear to be a reef development in sediments of Caratas age. Other *morros* such as those of Batatal, San Francisco de Macaira, and the large *morros* of San Juan are closely associated with rudistids and *Exogyra* or with shales carrying a Vidoño foram fauna and may be of quite different age—possibly as old as Middle Cretaceous.

## Eocene-Oligocene Strata

### *Merecure Group*

(Fig. 5; Pl. 9)

*Description.*—The Merecure formation was originally named from Quebrada Merecure, a small dry tributary of the Río Querecual (Hedberg, 1937a, p. 2000). However, much more nearly complete sections farther west on Río Capiricual and in the vicinity of Barcelona are considered as standard reference sections (Hedberg and Pyre, 1944, p. 15), and the formation has now been raised to the rank of a group.

The group is characterized principally by sandstones although shale probably makes up most of its thickness. Coal beds in the middle and upper part and orbitoid limestones at the base are also distinctive features. Total thickness in northeastern Anzóategui is 5500–7500 feet. The group is known in outcrop along the southern edge of the Serranía del Interior of Eastern Venezuela from Barcelona as far east as the vicinity of Aragua de Maturín. Beyond this point it is overlapped by younger Tertiary beds. It is not exposed between Barcelona and Píritu but otherwise crops out more or less continuously along the northern mountain front of northwestern Anzoátegui and Guárico. The group may be divided into several formations.

*Tinajitas formation.*—In the Barcelona area the base of the Merecure group is characterized by a persistent horizon of laterally discontinuous orbitoid-algal reef limestones which unconformably overlie the Santa Anita formation. A typical section is exposed on Quebrada Tinajitas just east of Barcelona. Here the limestones are overlain by some 600 feet of interbedded lavender-gray quartzitic sandstone, gray and greenish-gray slightly glauconitic sandstone and siltstone, and gray silty and locally foraminiferal shale which also are best included in the Tinajitas formation. In the Barcelona area the limestones are frequently developed only as isolated masses, rounded blocks, lenticles, or "boulders," apparently representing deposits formed at the borders of a limestone reef (Pl. 9, fig. 1). These grade laterally into more continuous limestone beds.

The formation in large part grades laterally east-southeastward along the mountain front into the Los Jabillos sandstone. However, even where the Los Jabillos sandstone is well developed there is usually between it and the typical Caratas formation sediments a small interval of calcareous sediments (grits, fine-grained sandstones, siltstones, and thin nodular limestones) with orbitoids, nummulites, calcareous algae, *Tubulostium*, *Oligopygus*, etc., to

which the term Tinajitas formation may still be applied. Such intervals ranging from a few feet to more than 100 feet in thickness are known on Río Querecual, Quebrada Pégua, Río Areo, Río de Oro, and elsewhere, and appear to form a transition zone from the Santa Anita to the Merecure (Pl. 9, fig. 2). West of Barcelona between Píritu and Río Unare orbitoid limestones and calcareous glauconitic sandstones of the Tinajitas formation are prominently exposed on Cerro San Antonio, Peñas Blancas, and other hills. The limestone here may reach 300 feet thick.

In northwestern Anzoátegui and northern Guárico the Tinajitas formation is represented at various scattered localities along the mountain front by thin orbitoid-algal limestones associated with conglomerates and glauconitic sandstones. The member is usually found in contact with Guayuta group sediments, and the conglomerates carry pebbles of chert, igneous rock, quartz, and feldspar.

*Los Jabillos sandstone* (Pl. 9, fig. 3).—East of Barcelona the basal portion of the Merecure group consists dominantly of very hard light-gray to pinkish-gray, thick-bedded, medium- to coarse-grained (and even pebbly) quartzitic sandstone 400 to 750 feet thick. This sandstone unit is known as the Los Jabillos sandstone, named from Cerro Los Jabillos just east of Río Querecual, and either immediately overlies or laterally replaces the Tinajitas formation. The two formations interfinger in the Naricual area. In many sections farther east thin shaly or calcareous intervals at the base or near the top of the Los Jabillos formation carry fossils. Such beds at the base of the Los Jabillos may be considered as belonging to the Tinajitas formation facies. Glauconite and fossil beds become increasingly common as the Los Jabillos formation is traced eastward into Monagas. The sandstone is very resistant and forms a prominent topographic marker across most of northeastern Anzoátegui until it grades westward into the less competent Tinajitas formation.

*Naricual formation.*—The thick upper part of the Merecure group above the Los Jabillos and Tinajitas formations in northeastern Anzoátegui is known as the Naricual formation. The type locality is in the vicinity of the coal mining town of Naricual, and the type section is well exposed in the valley of Río Naricual, in the mines, and along the roads of this area. The lower 1600 feet of the formation consists of carbonaceous shales, sandy shales, and sandstones, followed by the main coal-bearing series about 3000–3500 feet thick. This interval also includes numerous sandstones and shales. The upper part of the formation is about 1500 feet thick. It contains few coals and consists largely of a number of topographically prominent sandstones separated by barren gray shales. The lower more shaly part of the Naricual formation commonly forms a broad valley along the mountain front between the resistant Los Jabillos sandstone and the upper Naricual sand stones. The sandstones of the Naricual formation are typically massive, light-colored, quartzose, sugary to quartzitic, medium- to coarse-grained. The formation becomes more marine to the east. In the Barcelona area the formation is potentially an important source of coal.

*Areo shale.*—The lower portion of the Naricual formation grades eastward into a shaly glauconitic marine fossiliferous facies which is exposed in Río Areo, Río de Oro, and a few other streams in northwesternmost Monagas and northeasternmost Anzóategui. This unit has been variously called the Areo shale, the Juan Rosario shale, Buena Vista shale, and La Moya formation, but in the absence of a formal definition it is here referred to provisionally as the Areo shale from exposures on that river and its tributaries. The formation concists dominantly of (1) gray shales with seams of yellowish or reddish glauconitic ironstone concretions, and (2) occasional hard whitish gray quartzitic sandstones 5 to 30 feet thick characteristically fractured normal to the bedding. Molluscan fossils are commonly associated with the ironstone layers in the shale. There are some thin hard limestones. On Río Areo the unit appears to be more than 1000 feet thick. Molluscan fossils suggest Oligocene or even Miocene age, but foraminifera show close relation to those of the Tinajitas formation and include rare specimens of *Discocyclina* and *Lepidocyclina* suggesting uppermost Eocene age. The type section may be designated as on Río Areo east of the town of San Juan, from the mouth of Quebrada Juan Rosario upstream to the axis of the syncline just below El Salto.

This section does not expose the upper limit of the formation.

*Periquito formation.*—Wells of the Santa Ana-San Joaquín-Guario-Santa Rosa oil fields (Anaco fields) of central Anzoátegui have penetrated as much as 1900 feet into a formation which underlies the Oligo-Miocene Oficina formation and consists predominantly of light- to dark-gray, massive to poorly bedded, hard-fine- to coarse-grained sandstones and grits (Funkhouser *et al.*, 1948, p. 1870–1872). The continuity of the sandstones is broken by laminae and thin beds of hard black carbonaceous shale and intervals of gray claystone and siltstone, but sandstone makes up approximately 50 per cent of the formation. Cross-bedding is common, and secondary quartz growth is characteristic. Great lateral variability is indicated by the difficulties of electric-log correlation. This unit is believed to belong to the Merecure group and is probably equivalent to the Naricual formation but because of the greater sand content and the lack of coals is given a separate name derived from a locality in the vicinity of the Santa Ana field.

*Caño Dulce formation.*—Westward from Barcelona strata of the Naricual formation are covered until near Río Unare where they appear in a somewhat different facies known as Caño Dulce beds named from Quebrada Caño Dulce on the north flank of the chain of hills between Píritu and Río Unare. The formation consists of hard to friable, massive to well-bedded, whitish-gray to yellowish or dirty-brown sandstones and pebble conglomerates. Pebbles of gray or brownish chert predominate. There are some poorly exposed sandy carbonaceous shales and a few thin coal beds.

West of Río Unare the Caño Dulce beds continue in prominent development, and sandstone and conglomerate of this unit hold up the Fila Maestra ridge north of Guanape. In northeastern Guárico a thick interval of gray and black shale is developed in the midst of the formation. Coal beds are known at Guanape, Batatal, and near San Francisco and have been worked at Sabana Grande. Conglomerates are common. Some mottled claystones appear between the conglomerates and sandstones of the upper part of the formation. Still farther west beyond Altagracia the Caño Dulce beds constitute the "galera" sandstones near the base of Kaye's (1942) Guarumen group. The Caño Dulce beds probably are equivalent not only to the Naricual formation of the Merecure group but also to a part of the overlying Santa Inés group.

The Merecure group is generally unconformable on older formations along the northern edge of the Eastern Venezuela Basin. In the Barcelona area it rests on the weathered lower part of the Caratas formation. In northwestern Anzoátegui it rests on Vidoño shale, and in northern Guárico it is frequently found in contact with Guayuta sediments. Eastward from Barcelona the indication of unconformity becomes less distinct, and near the Anzoátegui-Monagas boundary there appears to be an unbroken transition from Santa Anita to Merecure groups. Southward from this area of uninterrupted deposition there is reason to believe that the unconformity develops rapidly.

The Merecure group everywhere appears to grade upward conformably into the overlying Santa Inés group. In northeastern Anzoátegui the coal-bearing deposits of the Naricual member give way upward to more marine sediments of the Santa Inés group. Black chert appears as a prominent detrital element in conglomerates and sandstones, and the typical light-colored massive quartzose Merecure sandstones (derived from the Guayana shield) are replaced by the characteristic brown, greenish-brownish-gray, and pepper and salt impure graywacke sandstones of the Santa Inés formations (derived from the borderland to the north and west). Because of the transitional relation the exact contact must be placed rather arbitrarily, but the general distinction between the two groups is clear (Hedberg and Pyre, 1944, p. 18). Likewise in the oil fields of central Anzoátegui the general distinction between the Periquito formation (Merecure) and the overlying Oficina formation (Santa Inés) is clear although no unconformity intervenes and the upper units of one may grade laterally into the basal part of the other.

In northwestern Anzoátegui and northern Guárico the relation between Merecure and basal Santa Inés formations is not only gradational, but the differences noted farther east become indistinct so that the Caño Dulce beds may well include both the Naricual formation equivalents and sandstones and conglomerates equivalent to the basal part of the Santa Inés. Kaye's (1942) Guarumen group west of Alta-

gracia includes both Merecure and lower Santa Inés.

The basal Merecure sediments represent the deposits of a shallow sea which in latest Eocene time transgressed eroded Cretaceous strata throughout most of the area of the present northern edge of the Eastern Venezuela Basin. The main Merecure seaway probably lay north of this line and opened to the east, and the Tinajitas limestone reefs were built up very near the shore line. Likewise the clean coarse sands of the Los Jabillos formation were built up as a shoreline deposit by rivers flowing northward from the Guayana shield. The sea was soon driven back from the present outcrop area first in the west and then in the east, and the deltaic Caño Dulce sandstones and conglomerates and the Naricual alternating sandstones and coal-swamp deposits soon replaced the initial marine sediments. The Periquito formation farther south represents a more sandy facies of the Naricual member deposited nearer the southern source and of more clearly fluvial or landward-delta origin.

Fossil data for the Merecure group of northeastern Anzoátegui has been summarized previously (Hedberg and Pyre, 1944, p. 18–21). The Tinajitas horizon with its distinctive assemblage of orbitoids, nummulites, Algae, *Tubulostium*, *Oligopygus*, and other forms has since been traced westward through Guárico and has even become established as a distinctive upper Eocene horizon across all Venezuela. The overlying Naricual, Caño Dulce, and Periquito formations are obviously younger and very probably extend well into the Oligocene.

The Merecure group correlates approximately with the lower part of the Guarumen group of western Guárico, the Paují shale of Barinas and the Maracaibo basin, the Cerro Misión and Cerro Campana beds of eastern Falcón, the Agua Negra series of Falcón-Lara, and the San Fernando formation of Soldado rock and Trinidad.

## Oligocene-Miocene Strata

### *Santa Inés Group*

(Figs. 5, 6; Pl. 10)

*Description.*—Following Mercure time the present area of the Eastern Venezuela Basin became the site of deposition of an immense thickness and volume of Oligocene and Miocene sediments. Near the axis of the depositional trough at the present northern edge of the Eastern Venezuela Basin some 30,000 feet of sediments was deposited. The geosynclinal basin opened eastward to the sea, and in Anzoátegui and Monagas sediments show a general eastward gradation of environments from nonmarine to brackish-water to shallow-water marine to deeper-water marine. In the case of previously discussed stratigraphic units usually only a narrow belt of outcropping strata along the present northern mountain front is available for study. In the case of the Oligocene-Miocene sediments, however, broad areas across the whole basin are exposed. Moreover, the extensive drilling for oil of the last 10 years has been concentrated in sediments of this age and has added tremendously to our knowledge. Consequently the facies variations of these sediments are much more fully known than are those of older sediments, and it is possible and desirable to classify them much more thoroughly into stratigraphic units on the basis of both vertical and lateral variations in lithology and lithogenetic character.

Sediments of the Oligocene-Miocene cycle of deposition may be conveniently assigned to one general group, here called the *Santa Inés group*. This group includes many laterally and vertically transitional subordinate stratigraphic units, more or less clearly defined by lithologic or lithogenetic characters, which are here ranked as formations. Boundaries between these units are naturally rather indefinite and subjective. The writer has attempted here to characterize briefly some of what seem to him to be the more important and better-established units, but he has little thought that these will all satisfy other workers, and in many cases were it not for the weight of earlier usage he might be tempted to make different subdivisions now that the over-all picture of these sediments is clearer. For convenience in the following résumé the outcrop areas of the various Santa Inés formations have been roughly outlined on the accompanying sketch map (Fig. 6). However, this gives only a part of the picture since much of the information on the stratigraphy of this group is based on well sections. The approximate relative strati-

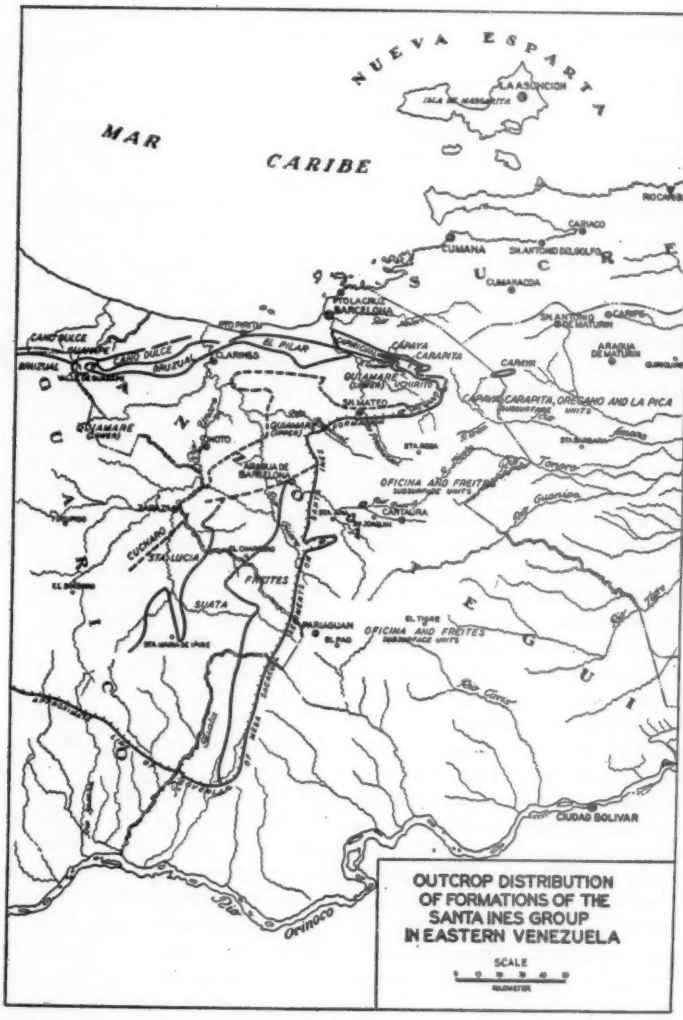


FIGURE 6.—OUTCROP DISTRIBUTION OF FORMATIONS OF THE SANTA INÉS GROUP IN EASTERN VENEZUELA

graphic position of individual formations is shown on the correlation chart of Figure 5.

The Santa Inés group appears to be essentially conformable and transitional with the Merecure group wherever it overlies that unit. However, on the southern limb of the basin it overlaps the Merecure and rests unconformably on Cretaceous formations or basement rock. A number of fairly extensive unconformities are developed within the Santa Inés group. Toward the east-central part of the basin the Santa Inés sediments usually grade upward conformably into the Sacacual group. Elsewhere they are generally unconformable below Sacacual sediments where these are present.

The different facies of the Santa Inés group afford fossils representing a great variety of environments. Small Foraminifera, nummulites,

orbitoids (*Miogypsina*), mollusks, echinoids, and vertebrates have been most useful in correlation and age determinations. The group clearly includes sediments from middle Oligocene to middle Miocene age and may include some lower Oligocene and some upper Miocene. It correlates approximately with the Agua Salada group of eastern Falcón; the group including the San Luis, Agua Clara, Cerro Pelado, Socorro, and Caujarao formations of western Falcón; and the combined Brasso and Springvale formations of northern Trinidad. Within this continuous series of sediments the Oligocene-Miocene boundary would be placed quite variably depending on the type of fossils used and on assumptions regarding the scope of each epoch.

*Carapita formation* (Pl. 10, fig. 2).—The type section of this formation and its faunal character have been discussed in some detail (Hedberg, 1937a, p. 2004–2008; 1937b; Hedberg and Pyre, 1944, p. 22). The formation is typically a highly foraminiferal moderately deep water marine gray shale with only rare sandstone beds. It is several thousand feet thick (perhaps as much as 7000 feet) where exposed in northeastern Anzoátegui. From its type section on Quebrada Carapita it pinches out rapidly westward grading into a shallower-water sandstone and shale facies (Capiricual formation). Eastward on the other hand the Carapita shale thickens at the expense of other members of the Santa Inés group, and subsurface data in northern Monagas indicate that here it may constitute the bulk of the total thickness of the group. The Carapita shale body is relatively incompetent and is frequently highly deformed and contorted as compared with adjacent more resistant units. (The type section on Quebrada Carapita is itself structurally deformed but nevertheless is the best single section of the formation.) The Carapita formation grades southward into the Oficina-Freites formations.

*Capiricual formation.*—The type section of this formation is on the Puerto La Cruz-Oficina highway (Hedberg and Pyre, 1944, p. 23) where it constitutes the basal 12,000 feet of the Santa Inés group. It is composed of dominantly shallow-water marine interbedded shales, sandstones, and conglomerates. There are a number of lignites and mottled claystones, largely in the lower part of the formation. The sandstones are generally brownish gray, greenish gray, brownish greenish gray, and pepper and salt color. They are commonly finely laminated Texture is typically fine- to medium-grained but abruptly defined streaks and lenses of black chert conglomerate occur throughout the whole formation and become more common westward. The sandstone units commonly range from 10 to 30 feet thick. Probably 80 per cent of the total thickness of the member is gray or brownish-gray shale and sandy shale. Most of the shales carry shallow-water Foraminifera. Seams of clay ironstone are common. To the west the formation disappears under the coastal plain south and west of Barcelona. To the east the Carapita shale is developed in its middle portion and rapidly replaces it except for a basal eastward-projecting tongue (Capaya tongue) and an overlying eastward-projecting tongue (Uchirito tongue).

*Capaya tongue or formation* (Pl. 10, fig. 1).—This unit is named from the community of San Antonio de Capaya about 6 kilometers south of Bergantín (Hedberg and Pyre, 1944, p. 22). It is an eastward-projecting tongue of the Capiricual formation which underlies the Carapita shale. It is about 2200 feet thick on Quebrada Carapita and in lithologic character resembles the Capiricual formation. It consists of fine-grained laminated gray, greenish-gray, brownish-greenish-gray, brownish, and pepper and salt sandstones; sandy or silty shales with thin seams of clay-ironstone concretions; and occasional coal beds. Fossils are scarce and largely limited to brackish-water Foraminifera. The formation is gradational into the Carapita shale above and into the Naricual formation of the Merecure group below, and exact definition is often difficult. It is included in the Merecure group by some and considered as a lower sandy phase of the Carapita formation by others. A small outlier of this formation is exposed on Río Areo some 50 kilometers east of the type area and is the easternmost exposure of any part of the Santa Inés group.

*Uchirito tongue or formation.*—The Uchirito formation includes the ridge-forming chert conglomerates and calcareously cemented sandstones immediately overlying the Carapita shale in its outcrop area in northeastern Anzoátegui and in part forming the topographically prominent Fila de Uchirito (Hedberg and Pyre, 1944,

p. 22). Between more resistant beds are gray and greenish-gray sandy and silty shales which yield a brackish-water to shallow marine foraminiferal fauna. The formation is about 4500 feet thick on Quebrada Carapita but thickens rapidly westward at the expense of the Carapita shale until it merges with the Capaya tongue to form the Capiricual formation. The uppermost part of the Uchirito formation in its type section is probably the lateral shallow-water equivalent of the nonmarine or brackish basal part of the Quiamare formation farther west. To the east the Uchirito formation becomes less conglomeratic and may eventually grade into a portion of the shale section assigned to the Carapita formation in wells of north-central Monagas.

*Quiamare formation.*—The type section of the lower part of this formation is exposed along the Puerto La Cruz-Oficina highway from the crest of Boca Tigre hill southward nearly to San Mateo (Hedberg and Pyre, 1944, p. 23). Its thickness is not accurately known but probably amounts to 10,000 feet. It consists largely of a series of gray, greenish, red and gray, brown and gray, and blue and gray claystones with some pebbly sandstones, carbonaceous shales, and lignites. It is largely nonmarine in the type section although there are occasional thin marine fossiliferous horizons. This part of the formation (lower member) forms the surface rock over the large Quiamare anticline. East of the type section the unit apparently grades laterally into shallow-water marine sediments to which the term Orégano facies or formation has been applied. West of the type section, in northwestern Anzoátegui and northeastern Guárico, the lower member of the Quiamare formation outcrops over a broad area of the rolling plains or llanos country. Here it is dominantly of continental and brackish-water origin, but there are occasional shallow-water marine horizons. Sandstones, conglomerates, and thin limestones form most of the outcrops, but actually claystones and shales predominate in the formation. Common lithologic types are highly lenticular coarse to fine, greenish-gray, light-greenish, or greenish-brown thin-bedded to massive sandstones; distinctive clay-pebble conglomerates; thin dense cream to brown locally fossiliferous limestones and calcareous siltstones ("Onoto limestone"); and red and gray mottled claystones. Brackish-water to shallow marine mollusks are fairly common as are also vertebrate remains.

The upper member of the Quiamare formation includes the topographically prominent Dividive ridge running eastward from San Mateo and also includes still younger sediments just north of this ridge and also occupying the synclinal area west of San Mateo and north of Aragua de Barcelona. It consists of an alternation of fresh-water, brackish-water, and shallow marine sediments. Sandy limestones and calcareous sandstones are responsible for the topography of the Dividive ridge. Common lithologic types in the llanos area farther west are red to purplish-weathering claystones, brown sandstones, and occasional lignites and sandy limestones. Brackish- and shallow-water marine mollusks occur at a number of horizons. Low dips and low relief make it difficult to piece together a comprehensive section and the gradual lateral changes which take place in lithology in this western area makes it difficult to draw any definite boundary between upper and lower Quiamare members and between the Quiamare sediments in general and the Suata, Santa Lucía, and Cucharo formation facies farther southwest.

*"Orégano formation."*—This still rather ill-defined unit represents the shallow-water marine facies into which the Uchirito and Quiamare formations apparently grade in an easterly direction, (The term is used here provisionally pending further clarification of stratigraphic relations in this part of the section of northeastern Anzoátegui and northwestern Monagas.) It is partially exposed on Río Orégano above the Carapita shale and also forms the easternmost outcropping sediments of the Santa Inés group between Río Orégano and Cerro Pelado. Part of the formation is very well exposed in the big bend of Río Orégano about a kilometer south of the Santa Inés-Urica road crossing. The section here consists largely of bluish-gray shale and massive buff sandstone. Echinoids, mollusks, and crabs are common in the shale. Although outcrops are poor, well sections farther east suggest an extensive subsurface unit consisting of shallow-water marine fossiliferous shales, sandstones and conglomerates constituting the transition, zone between the dominantly nonmarine or

brackish-water Santa Inés sediments to the west and the moderately deep water marine foraminiferal Santa Inés shales to the east. Locally in uplift areas along the northern mountain front well sections indicate that sediments of Orégano facies overlap older pre-Santa Inés rocks.

*El Pilar conglomerate.*—In the district of Bolívar in northern Anzoátegui between Barcelona and Píritu a tremendously thick conglomerate series develops in the lower part of the Quiamare formation. Estimates of the thickness of this conglomerate run as high as 15,000 feet. The thickness, areal extent, and distinctive lithologic character of this facies warrant ranking it as a formation. It is commonly referred to by the name of El Pilar conglomerate from the town of El Pilar near the southern edge of its outcrop. The formation consists of beds, commonly ranging from 20 to 80 feet thick of poorly cemented pebble, cobble, and boulder conglomerate usually separated by brown and red or brown and bluish-gray silty claystones. Boulders are predominantly of black chert with a minor amount of bluish-gray limestone. There are surprisingly few beds of sandstone in the formation, and the change in grade size from conglomerate to claystone is very abrupt. Some conglomeratic or sandy limestones with fossil oysters are known. The formation is topographically prominent and forms the San Bernardino hills between Barcelona and Píritu. However, because of the poorly cemented character of the conglomerates and their tendency to break down into boulder-strewn areas, outcrops of the formation are not very good. The base of the formation is not known since it is overlapped on the north by recent coastal alluvium. Upward and eastward it grades into sediments of the Quiamare formation.

*Bruzual claystone.*—From the vicinity of Río Unare westward to Altagracia de Orituco a large area of northeastern Anzoátegui and northwestern Guárico adjacent to the northern mountain front is covered by mottled claystones. This formation is here designated the *Bruzual claystone* from the District of Bruzual, State of Anzoátegui, where it is typically developed. The claystones are characteristically mottled in dull shades of red, brown, purple, greenish gray, and gray. They commonly weather to a tan color with bands of red. There are some interbedded friable gravelly conglomerates and thin gray to purple fine- to coarse-grained sandstones. Lack of continuous exposures makes it difficult to designate any one satisfactory type section. The formation is probably at least 5000 feet thick, though accurate figures are lacking. Because of the absence of resistant beds the outcrop area is one of low nearly featureless relief. The only fossils found are rare brackish-water Foraminifera, and the formation is believed to be dominantly a fresh- to brackish-water deposit. It is laterally equivalent to some part of the Capiricual formation and perhaps occupies approximately the same stratigraphic horizons as the western end of the Carapita shale.

*Guanape conglomerate.*—In the vicinity of Valle de Guanape a boulder conglomerate several hundred feet thick and forming steep walls and prominent topography overlies the Bruzual clays and apparently forms a part of the Santa Inés group. Boulders are largely chert, though some are sandstone and quartz. Its lithologic character and stratigraphic position above the Bruzual claystones suggest its approximate equivalence in age to the El Pilar conglomerate.

*Caño Dulce formation.*—This unit, which was discussed under Merecure group, probably also includes some sediments equivalent to the basal part of the Capiricual formation of the Santa Inés group.

*Peña Mota conglomerate.*—A thick series of coarse reddish, purplish, or greenish pebble to boulder conglomerates interbedded with dull red, purple, brown, and green mottled sandy claystone occurs at the town of Altagracia de Orituco in Guárico and has been named from the Peña Mota ridge at the north end of the town. Pebbles and cobbles are largely greenish sandstone, quartz, and chert. These beds have been noted at several points along the northern Guárico mountain front at the top of the Caño Dulce beds and below the Bruzual claystones. They appear to represent fluvial or deltaic boulder deposits dumped into the basin from a closely adjacent land mass to the north.

*Chaguaramas formation.*—This term has been applied to far western equivalents of portions of the Oficina and Quiamare formations, outcropping between Altagracia de Orituco, Chaguaramas, and Las Mercedes in the State of Guárico. The limits of these beds have not been

formally defined. They consist of brackish-water shales and sandstones and fresh-water claystones. Coarse-shelled mollusk fragments and vertebrate remains are known. The distinctive clay-pebble conglomerates of the Quiamare formation are well developed in the Chaguaramas beds also.

*"Roblecitos shale."*—This is a subsurface unit which underlies the Chaguaramas formation in Guárico. It consists in eastern Guárico of some 2000 feet of dark-gray shale with a few thin siltstones and sandstones. The unit here is of marine to brackish-water origin and contains a fairly abundant foraminiferal fauna related to that of the Carapita formation. It is probably the same age as the Bruzual claystone and may also be the western time equivalent of part of the Carapita shale, perhaps representing a similar but independent deep-water depositional basin. The "Roblecitos shale" is underlain by the "La Pascua sandstone" which is probably equivalent to the Capaya formation and the Merecure group.

*Guarumen group.*—In western Guárico near Ortiz, Kaye (1942) has used the term Guarumen group to include the series of sandstones with intervening shales and siltstones which constitute the topographically prominent "galera" ridges of this region. This group appears to include at least 9000 feet of section and is probably equivalent in age to the Merecure group and the lower part of the Santa Inés group farther east.

*Suata formation* (Pl. 10, fig. 3).—This formation has been named by geologists of the Creole Petroleum Corporation from the town of Suata in the District of Monagas, State of Anzoátegui. In the type area the upper part of the formation as seen in outcrop consists of some 250 feet of finely interlaminated sandstone and clay-shale, variously colored claystones and sandy claystones, and thin lignites and lignitic shales. The lower part consists of about 400 feet of gray-green to yellowish-green claystones; siltstones; and fine sandstones with some thin lignites. Concretions are common. The formation occupies a broad gently dipping northeast-southwest trending belt in western Anzoátegui and southeastern Guárico. It grades northward into the upper middle part of the Quiamare formation. The formation is laterally variable and includes several mild disconformities. The sediments are dominantly of brackish-water origin, and brackish-water mollusks and plant fossils are fairly common. Low dips and lack of resistant beds give a gently rolling landscape typical of the Venezuelan llanos.

*Santa Lucía formation.*—This formation which underlies the Suata formation is named from Quebrada Santa Lucía in southeastern Guárico. It consists of several hundred feet of alternating greenish and gray claystones and sandy claystones, mottled claystones, carbonaceous shales, and lignite beds. Brackish-water mollusks occur at a few horizons. The formation is dominantly a coastal-swamp deposit, and coal beds are characteristic.

*Cucharo formation.*—This name has been applied by geologists of the Creole Petroleum Corporation to a "littoral" shallow marine to brackish facies of yellowish cross-bedded sandstones, greenish calcareous claystones and sandy claystones, and red- or purple-weathering claystones. The name is derived from Quebrada Cucharo, a tributary of Quebrada Honda northeast of the town of Socorro, in the State of Guárico. The Cucharo formation in general underlies the Santa Lucía formation but is closely related to it; the boundary between the two is indistinct. It grades northward into the Quiamare formation from which there is also no very distinct line of separation.

*Oficina formation.*—The term Oficina formation is applied to the lower and major part of the Oligocene-Miocene section encountered in the wells of the Oficina field and associated fields in south-central Anzoátegui (Hedberg, *et al.*, 1947). This section consists dominantly of alternations of gray, dark-gray, and brownish shales, interlaminated and interbedded with light-gray fine- to coarse-grained sandstones and siltstones. Important minor constituents are thin lignites and lignitic shales, green and light-gray claystones with siderite spherules, sideritic glauconitic sandstones, and thin cone-in-cone limestones. Carbonaceous matter is common throughout, and individual well sections may contain as many as 40 or 50 lignite beds, reaching up to 2 feet in thickness. Many of the sandstones may be called grits, and some are conglomeratic with quartz and chert pebbles. Sandstones in general are thicker, coarser, and more frequent toward the base of the formation. Total thickness of the unit in the Greater

Oficina area ranges from about 1400 to about 4000 feet. The formation represents a repeated alternation of shallow-water marine, brackish-water, and coal-swamp environments.

The Oficina formation has been traced eastward through wells in southern Anzoátegui and Monagas to the Territory of Delta Amacuro. It is also present in the Anaco group of oil fields northwest of Oficina, where it has thickened to 7500–10,000 feet.[4] The Oficina facies of the Santa Inés group also undoubtedly extends far northward and eastward within the Eastern Venezuela Basin, and it can be correlated quite definitely with a portion of the Brasso group of sediments in the northern basin of Trinidad. The formation name has been applied only to subsurface developments although the time equivalence of the Oficina formation to the outcropping Suata and older Santa Inés group units farther west has been definitely established. (An exception to this is the few small outcrops on the Santa Ana and San Joaquín structures where the term Oficina formation is applied).

*Freites formation.*—Wells drilled for oil in the central part of the District of Freites, State of Anzoátegui, showed a distinctive Middle Miocene stratigraphic unit characterized by green to greenish-gray clay-shales resting conformably on the Oficina formation (Hedberg *et al.*, 1947). This was called Freites formation from the name of the district. The formation is about 1600 feet thick in the Oficina field but ranges elsewhere from 1000 to 2500 feet. In the Oficina area it is divided into three members. The lower member consists of greenish shales and yellowish-green medium- to coarse-grained glauconitic highly fossiliferous marine sandstones with a few thin sandy limestones. The middle member is purely shale and is of marine to brackish-water origin. The upper member in addition to the characteristic greenish shales carries a few beds of light-gray slightly glauconitic sandstone. The upper member is of brackish-water origin.

As evidenced by excellent well-section correlations, the Freites formation extends with remarkable lithological constancy throughout the southern and south-central part of both Anzoátegui and Monagas, and it has also been identified in the north-central part of the Eastern Venezuela Basin. Toward the west it has been traced by core drilling up dip to the surface outcrop, and it constitutes the uppermost exposed unit of the Santa Inés group in central and southern Anzoátegui, where it forms a broad band of outcrop from the vicinity of Aragua de Barcelona on the north southward into the district of Monagas, Anzoátegui. In the central part of this outcrop belt, it possesses much the same character as in the Oficina field. To the south it shows a more brackish-water environment and to the north it becomes sandy and less persistently marine and grades laterally into the upper member of the Quiamare formation. The shaly nature of the formation is reflected in the featureless topography of its outcrop belt and the characteristic growth of savanna palms. The outcropping unit has been referred to as Aragua formation and as Punche formation, but because of the widespread commercial usage of the term Freites formation it is believed desirable to accept this name for the outcropping unit also.

### *La Pica Formation*

The type section of the La Pica formation is the interval 3100–5857 feet in the La Pica No. 1 well, District of Maturín, State of Monagas. The formation consists generally of gray brackish to marine shales and silty shales with important local developments of fine-grained clayey sandstones. Maximum thickness may be considerably greater than that of the type section. The formation is characterized by a meager but very distinct fauna of arenaceous, and dwarfed brownish calcareous Foraminifera. It is known principally from well sections in northern Monagas and Delta Amacuro and does not appear anywhere in outcrop. It may be correlated with the Forest and Cruse formations of Trinidad on the basis of the distinctive foraminiferal fauna. It is probably in part approximately equivalent in age to the Freites formation and other members of the upper part of the Santa Inés group, but represents a much different facies. It also appears to be laterally gradational in part into the Las Piedras forma-

[4] Recent drilling indicates that the basal sandstones of the Oficina formation in the Greater Oficina fields are *lithologically* continuous with the uppermost Periquito formation of the Anaco fields although the former are probably somewhat younger.

tion of the overlying Sacacual group. In the Santa Barbara area of northeastern Monagas the La Pica formation rests with strong angular unconformity on the Carapita shale. The formation is probably of late Middle Miocene and late Miocene age.

The La Pica formation is here considered apart from either the Santa Inés group or the Sacacual group although it is probably the partial time equivalent of a portion of both groups.

## Miocene-Pliocene Strata

### *Sacacual Group*

(Fig. 5; Pl. 11, fig. 1)

*Description.*—Sediments of probable Pliocene and Late Miocene age overlie the Santa Inés group throughout the eastern portion of the Eastern Venezuela Basin and locally overlap unconformably on older rocks on both its northern and southern edges. These sediments are dominantly of continental or brackish-water origin and were deposited within the brackish-water gulf extending westward from the Atlantic at this time or under fresh-water conditions in lakes and rivers near its margins. They comprise several different facies totalling several thousand feet, and a number of formation names are in use. However, this whole group of related sediments, above the uppermost marine beds of the Santa Inés group (and La Pica formation) and below the Pleistocene Mesa formation, may conveniently be included under the term Sacacual group. The name is derived from Quebrada Sacacual, a tributary of Río Güere in north-central Anzoátegui which cuts through these sediments for 10 kilometers in the vicinity of the Puerto La Cruz highway. The boundaries of the Sacacual group essentially coincide with those of the Pliocene-Miocene area of Figure 4. Two principal formations of the several which have been recognized are discussed.

*Quiriquire formation.*—This formation is named from the Quiriquire field in northern Monagas where it includes as much as 5000 feet of section overlying the so-called Monagas shale (equivalent to Carapita shale (Regan, 1938). In the Quiriquire field and in general near the mountain front poorly consolidated sandstones with boulder, cobble, and pebble conglomerates unite with claystones to make up the bulk of the formation. Farther south in Monagas both wells and outcrops show the formation to consist dominantly of claystones and siltstones. These are commonly light gray, greenish gray, or yellowish brown, and locally are mottled with red. There are some poorly consolidated sandstones and conglomerates. The Quiriquire formation farther west in Anzoátequi consists characteristically of tan-brown claystones and siltstones and tan or red and light-gray mottled claystones. Quiriquire sediments are commonly calcareous. They are highly variable laterally.

The Quiriquire formation forms the dominant part of the Sacacual group adjacent to the northern mountain front and adjacent to the northwestern edge of the mesa in northeastern Anzoátegui. However, it grades laterally and downward into the Las Piedras formation of the Sacacual group and thins southward and eastward to the advantage of the Las Piedras formation. Outcrops are known as far south as the Tigre Rivér in Eastern Anzoátegui where however it is only a few hundred feet thick. The tough claystones and coarser clastics of the formation frequently form a hilly to semi-badlands topography of considerable relief. The formation is entirely of continental origin, and fossils are largely limited to vertebrate remains. A fossil turtle horizon is widespread in northeastern Anzoátegui.

*Las Piedras formation.*—The Las Piedras formation is named from the Las Piedras No. 1 well of the California Petroleum Corporation in the northern part of the District of Maturín, State of Monagas, where it includes the well section from about 1600 to about 4500 feet depth. It consists typically of light-gray and greenish-gray interlaminated fine-grained more or less carbonaceous sandstone and clay-shale, light-gray fine-grained friable sandstones, and lignites and lignitic shale. There are occasional beds of hard-green arenaceous limestone and gray sideritic claystone.

The formation is persistent in rather typical development over a broad area; it is known from well sections throughout Monagas, Delta Amacuro, and central and southern Anzoátegui. Its principal outcrop area is in central and southern Anzoátegui just west of the mesa front where it has been called Campo Santo formation. However, the subsurface name is

more widely used. The thickness of the formation ranges from about 1500 feet in the Anzoátegui outcrops to as much as 5000 feet in well sections in the eastern axial portion of the basin. A more sandy, less well-bedded, and more nearly fresh-water deposit at the top of the formation in south-central Anzoátegui is known as the *Algarrobo formation* from type exposures on Quebrada Algarrobo, a tributary of Río Pao. A unit of reddish sandstone equivalent to the upper part of the Sacacual group in the area east of the Leona oil field in south-central Anzoátegui is known as *Pando formation*.

The Las Piedras formation in part underlies the Quiriquire formation and is in part its lateral equivalent. It is distinguished from the Quiriquire formation principally by its bedded and laminated character as compared with the massive claystones and siltstones of the Quiriquire. The contact between the two is usually gradational.

The Las Piedras formation is dominantly a brackish-water deposit, and fossils are largely limited to fish remains and occasional fresh- to brackish-water mollusks and Foraminifera. Well-preserved plant remains have been found in various localities. Brackish-water fossils described from the area north of Caicara de Maturín (Palmer, 1945; Marshall, 1934) probably come from this formation. A Pliocene age is indicated.

Sediments of the Sacacual group are unconformable on the Santa Inés group of formations and on older rocks at the northern and extreme southern edges of the Eastern Venezuela Basin and in its western part. This unconformity is frequently distinctly angular. However, throughout much of the median portion of the basin there was apparently continuous deposition from Santa Inés to Sacacual (Las Piedras) time, and the change from one group to the other is marked principally by the upward disappearance of marine fossils and the change from a dominant marine environment to exclusively nonmarine conditions. Sediments of the Santa Inés group are also in general better lithified than those of the Sacacual group. Within the Sacacual group itself there are numerous local unconformities, as might be expected from its conditions of deposition.

The Sacacual group is considered uppermost Miocene and Pliocene largely because of its position stratigraphically above middle or upper Miocene strata of the Santa Inés group and below the Mesa formation of supposed Pleistocene age.

### Pleistocene Strata

#### *Mesa Formation*

(Fig. 4; Pl. 11)

This youngest stratigraphic unit is named from the flat topographic feature known throughout Eastern Venezuela as the "mesa" which is largely constituted by rocks of this formation. The formation covers broad areas particularly in eastern and southern Anzoátegui, southern Guárico, and Monagas, where it forms the distinctive savanna-covered mesas bordered by steep cliffs (*farallones*) or dissected badland areas.

No specific type section has been designated, but innumerable fairly complete reference sections are exposed along the mesa front. The formation of the interior plains consists typically of massive but cross-bedded pink, buff, or red friable sandstones with some pebble conglomerates and some gray and mottled claystones and sandy claystone lenses. The resistant "caprock" which holds up the mesas is a massive dark-red ferruginous pebbly and sandy rock composed mainly of residual gravels which have been indurated by ground-water cementation. It is usually immediately underlain by red and gray mottled sandy claystone. The Mesa formation is usually 50 to 250 feet thick but may be as much as 500 feet. It thickens northward toward the mountain front and also coarsens in this direction and to the west. Pebbles, cobbles, and boulders consist largely of quartz, quartzitic sandstone, and chert.

The Mesa formation is of diverse origins. Toward the north it is largely formed by the deposits of coalescing alluvial fans. Farther south it includes fluvial, eolian, and residual sediments and may also include playa and small lake deposits. It is everywhere of continental origin. Eastward in Territorio Delta Amacuro it grades into dominantly shaly and silty delta deposits to which the name *Paria formation* has been given.

The Mesa formation usually rests uncon-

formably on all older rocks, and this relation is quite clear along the mountain front and in areas of local uplift. Toward the center of the basin and especially to the east, however, the break between Sacacual and Mesa beds is obscure, and the contact appears to be transitional. In many cases the original Mesa deposit apparently was reworked by more recent eastward-flowing streams; where these streams cut down to the Sacacual group the unconformity is most pronounced. The Mesa formation in general possesses more gravel and fewer claystones than the Sacacual sediments and is less frequently calcareous.

Fossils in the Mesa formation are limited largely to local occurrences of petrified wood. The formation is commonly considered Pleistocene but without any very definite evidence. Certainly it is older than the last major uplift of the mountains since its surface shows gentle tilting and warping, and the formation is locally faulted. The remnants of Mesa formation in the foothills are definitely older than the alluvium of the present streams which have cut several hundred feet below the base of the Mesa beds.

## Geological History

The oldest records of the history of the Eastern Venezuela Basin undoubtedly lie in the basement rocks outcropping on the borders of the Guayana shield and encountered by wells drilled for petroleum along the southern limb of the basin. However, these have yielded little information except that sedimentary rocks of pre-Mesozoic age were deposited in this region, intruded and metamorphosed, and then worn away until only remnants of them remain as schists and phyllites in a dominantly igneous basement. No Paleozoic sediments have thus far been definitely identified in Eastern Venezuela although Ordovician, Devonian, Carboniferous, and Permo-Carboniferous strata occur in the Andes of Western Venezuela and it is therefore not improbable that Paleozoic sediments exist below the younger beds in the northern part of Eastern Venezuela.

Evidence has been given (Hedberg, 1942, p. 200–201) for considering the Hato Viejo and Carrizal formations equivalent to the La Quinta formation of Western Venezuela and thus possibly of Triassic-Jurassic age. If this should be correct it would appear that, following probable diastrophic movements near the close of the Paleozoic, the Eastern Venezuela region was essentially peneplained and was then invaded by an early Mesozoic sea. In southwestern Anzoátegui and in Guárico where the Hato Viejo and Carrizal formations are known sedimentation was largely in brackish water at the edge of a gulf or embayment extending into the continent from an open sea to the east or northeast. These sediments graded westward into the continental or deltaic deposits of the La Quinta formation and possibly southward into the fluvial and eolian deposits of the Roraima formation of the Guayana shield. It would appear reasonable that to the east and north they graded into more definitely marine strata. Uppermost Jurassic (Tithonian) marine fossils are known in the northern range of Trinidad in association with a thick phyllite series which may in part represent the metamorphosed equivalents of the Hato Viejo-Carrizal beds. Likewise a similar series of phyllites, with some limestone and chert, is known in the Serranía de la Costa of the Peninsulas of Paria and Araya in northeastern Venezuela, and these also may at least in part represent the metamorphosed marine equivalents of the Hato Viejo-Carrizal sequence. Finally it should be mentioned that supposed Jurassic fossils have been reported from limestones in a metamorphosed schist-limestone-quartzite series in the Serranía de la Costa east of Caracas (Wolcott, 1943).

During the deposition of the supposed early Mesozoic sediments of Venezuela vigorous igneous activity is indicated by tuffs, flows, and intrusions; the occurrence of biotite, widespread in these sediments though rare elsewhere, may represent an admixture of pyroclastic material.

Evidence of late Jurassic (Nevadan) orogeny in Venezuela has been presented previously (Hedberg, 1942, p. 205–206). Certainly in the southwestern part of the eastern Venezuela region there was pronounced folding, uplift, and erosion which resulted in the increased lithification shown by the Hato Viejo-Carrizal sediments and their separation from the Cretaceous Temblador formation by a marked angular unconformity. If some of the metamorphosed sediments of the Serranía de la Costa and the

Northern Range of Trinidad should be proved to be of early Mesozoic age the metamorphism of these rocks as compared with adjacent Cretaceous strata may very well be attributed to this orogeny.

Erosion following the folding and uplift of the Hato Viejo-Carrizal sediments appears to have left at least the southern portion of the Eastern Venezuela region essentially peneplained, with these supposed early Mesozoic sediments removed in southeastern Anzoátegui and southern Monagas and preserved only in a downwarped basin westward from the vicinity of Pariaguán. Farther north or northeastward toward the open sea the unconformity between early Mesozoic and Cretaceous sediments must naturally have diminished and disappeared, and it is even possible that the Tithonian sediments of Trinidad may grade upward without a break into Cretaceous strata.

In early Cretaceous time general subsidence of the Eastern Venezuelan region began again. Whether an actual geosyncline was in existence here at this time is uncertain. If so, its northern lay beyond the limits of the present land, and no evidence of it remains. The early Cretaceous sediments of the region may be considered equally well as deposits on the southern limb of such a geosyncline or as merely a continental-shelf deposit bordering the open ocean.

The base of the Barranquín formation is not exposed, and while the lower less marine part of the formation may include Neocomian strata (and may even be transitional with the Jurassic of Trinidad) the earliest dated marine deposits of this information are of Barremian-Aptian age. The early Cretaceous transgression was doubtless from the north or northeast, and hence the basal Barranquín is probably older than the Temblador formation of the southern limb of the present basin.

The earliest exposed sediments preserved by the subsidence of the Eastern Venezuela region in Cretaceous time are dominantly nonmarine fluvial or deltaic sands, mottled claystones, and carbonaceous coastal-swamp deposits representing sediments derived largely from the Guayana shield and deposited just south of and on the landward side of the advancing sea. These are followed by alternations of shallow-water marine beds with continental deposits and finally by persistent shallow-water marine strata. The advance of the sea was slow and marked by many fluctuations, and the border of permanent deposition was only gradually shifted southward. Thus while the lower Temblador formation of the southern part of the region most nearly resembles the lower Barranquín formation of the northern section in facies it is probably considerably younger and may be the *time* equivalent of the upper Barranquín, El Cantil, or Chimana formations.

In Aptian-Albian time, a relatively clear shallow sea covered the northern part of the region. Thick limestone "reefs" were formed by corals and other marine animals and plants (El Cantil formation) and were bordered by highly glauconitic clastic sediments of great lateral variability. In this northern area the rate of sedimentation appears to have become very slow in Middle Cretaceous time as indicated by the accumulation of beds of nearly pure glauconite and the presence of only a very thin interval of sediments occupying the interval between Middle Albian and Turonian strata.

A marked change in character of sedimentation occurred near the beginning of Turonian time when the deposition of the black foraminiferal bituminous limestones of the Guayuta group began. These strata must have been laid down in fairly deep water in a sea swarming with planktonic life but into which very little detrital clastic material was brought. Moreover, stagnant bottom conditions inhibited benthonic life and allowed the remains of planktonic Foraminifera to accumulate. The lack of bottom currents is indicated by the fine uniform lamination of these sediments produced by alternately rich and poor layers of Foraminifera. This period of Turonian-Senonian time probably marked the maximum depth and southward extent of the Cretaceous sea. The shoreward facies may be represented by some part of the Upper Temblador formation in the southern part of the basin. The evidence of restricted bottom circulation may indicate the initial development of the borderland to the north as a barrier to open-sea conditions although separated from the continental shield by deep water.

During the latter part of Guayuta time there was marked shallowing of the sea and an in-

crease in the amount of clastic material brought into the northern part of the region. The local abundance of chert in the San Antonio formation suggests a high content of silica in the waters flowing into the sea (or perhaps submarine volcanic activity) and also indicates conditions favorable for its precipitation. Emergence of border areas appears to have begun, and a change to a dominantly sandstone and shale facies westward along the Serranía del Interior into central Venezuela indicates that a land area was emerging there.

With gradual epeirogenic uplift, the whole Eastern Venezuela region was converted either to local land areas or to shallow seas. The thick San Juan sandstone and other Santa Anita sandstones were formed, while in more quiet areas, where less clastic sediment was being brought in, the Vidoño shale with its rich benthonic foraminiferal fauna was deposited. Part of the brackish-water to marine Upper Temblador formation with its glauconite beds, coarse sandstones, and thin limestones may represent the southern phase of this late Cretaceous deposition. Local uplift in Central Venezuela actually furnished coarse conglomerates at this time in the San Juan de los Morros area, while local basins farther east maintained Guayuta-like conditions in which pelagic Foraminifera accumulated in pockets of stagnant water.

The general conditons of the late Cretaceous continued into Paleocene time with an increasing tendency to emergence of the Cretaceous sea floor about the edges of the Guayana shield and in Central Venezuela. Even more pronounced uplift may have occurred north of the present coast line (Margarita, etc.), and the characteristic Paleocene limestone reefs of Soldado Rock, Cerro Corazón (near Urica), and some of the morros of north-central Venezuela were probably formed in a narrow shallow sea trough coinciding closely with the present mountain area of northeastern and north-central Venezuela and opening eastward. In early Eocene time all of Central Venezuela became emergent, but a shallow sea still persisted in the area of the Serranía del Interior of Eastern Venezuela. Static conditions prevailed, and the low-lying recently emerged adjacent land supplied relatively little detritus. The glauconitic calcareous and dolomitic siltstones of the Caratas formation are believed to have accumulated very slowly and to represent a relatively long period of Eocene time.

The Santa Anita group of sediments appears to indicate essentially continuous deposition from Cretaceous through Paleocene into Eocene time in *northeastern* Venezuela. Moreover in a very restricted area in northeastern Anzoátegui and northwestern Monagas the Caratas formation of this group appears to grade without a noticeable break upward into the Upper Eocene Mercure formations, thus apparently representing not only Paleocene but Lower and Middle Eocene deposition. However, the area of the Serranía del Interior where these observations are made represents only a very local area which probably lay near the center of the regional axis of subsidence. In general, the time from Late Cretaceous to Late Eocene was one of widespread emergence accompanied by tectonic movements and pronounced changes in paleogeography.

Away from the area of continuous deposition in northeastern Anzoátegui a major unconformity extends southward, westward, and probably northward with increasing time value toward the edges of the Eastern Venezuela region, separating Santa Anita and older Cretaceous strata below from the Upper Eocene, Oligocene, and younger strata of the Merecure-Santa Inés cycle of deposition above. The magnitude of this unconformity at any one place is a function of (1) the time of cessation of Cretaceous or Early Tertiary deposition, (2) the amount of erosion of Early Tertiary and/or Cretaceous deposits, and (3) the time of beginning of Merecure or Santa Inés deposition. The maximum extent of the unconformity appears to have been attained within Early or Middle Eocene time, and consequently as a whole it cannot be properly referred to as a Cretaceous-Tertiary unconformity although it frequently separates Cretaceous from Tertiary rocks.

Because the bulk of Eastern Venezuela lay on the relatively stable northern slope of the continent, sharp angular discordance at the unconformity is rare, and owing to local reworking and similarity of environment above and below, the exact position of the unconformity in this region is often difficult to determine. It is to be expected, however, that

farther north in the vicinity of the mobile borderland or farther west where regional uplift was greater the angular relation will be found to be more pronounced, and basal conglomerates and more abrupt contacts will exist. It is significant that in the Serranía del Interior of Eastern Venezuela there was at no time sufficient disturbance to create any strong angular relation between any of the several formations or sufficient erosion to break the normal sequence of formational units or even very markedly reduce individual formation thicknesses.

The date of origin of the Eastern Venezuela Geosyncline, a subsiding trough between the main Guayana continental mass and an outlying borderland to which the name Paria has been given, is somewhat problematical since we have no record of a northern shore to the Early and Middle Cretaceous sea and only weak inferential evidence of a northern borderland during Upper Cretaceous time. Margarita, however, furnishes definite evidence of at least a local land area to the north in Middle Eocene time, and from Late Eocene time onward the evidence for the Eastern Venezuela Geosyncline is reasonably clear. Communication with the open ocean was largely to the east. Southward pressure of the Parian borderland and consequent repeated buckling of the northern border of the sedimentary trough shifted the axis of the geosyncline progressively southward during Tertiary time (Hedberg, 1937a) (Fig. 2).

A new cycle of deposition began in Eastern Venezuela with the Merecure group. The main Merecure seaway corresponding to the axis of the geosyncline probably lay north of the present shore line, but an advance of this sea spread the shallow-water orbitoid-algal limestones of the Tinajitas formation far westward over the eroded surface of the Santa Anita and older Cretaceous sediments even to Western Venezuela, while northward-flowing rivers from the Guayana shield built up the clean coarse sands of the Los Jabillos formation as a shoreline deposit. Following this early marine invasion the sea withdrew, first in the west and then in the east, perhaps due to uplift in the surrounding areas which brought in clastic sediment at a rate exceeding that of the subsidence of the basin. An alternation of coarse sandstones with coal-swamp deposits (Naricual formation) replaced the marine sediments in part of northeastern Venezuela while farther west they were superseded by the deltaic Caño Dulce sandstones and conglomerates. However, in spite of the retreat of the sea the area of deposition was extended farther and farther southward onto the flank of the Guayana shield, and in central Anzoátegui the upper part of the formation is represented by the Periquito formation of fluvial or landward delta origin. In northern Monagas marine conditions persisted even during later Merecure time and resulted in the deposition of the dominantly marine Areo shale.

The latest Eocene or early Oligocene was a time of pronounced crustal movements over much of Venezuela. In western Venezuela the Venezuelan Andes were uplifted, separating the Maracaibo basin from the Apure basin; in Eastern Venezuela, similarly, folding and uplift appear to have taken place in the Parian borderland. At the same time there was pronounced subsidence of the Eastern Venezuela Geosyncline with the shifting of its axis southward to the latitude of the present Serranía del Interior in northeast Venezuela and somewhat farther south in north-central Venezuela. Up to this time about all of Eastern Venezuela had lain south of the axis of the geosyncline and had been the site of deposition of sediments derived largely from the Guayana shield. Now, however, in Middle Oligocene to Middle Miocene time coarse clastic sediments, derived in large part from erosion of Cretaceous rocks in the rapidly rising borderland of Paria, were dumped into the area of the present Serranía del Interior and were spread far southward over the flank of the Guayana shield to constitute the various facies of the Santa Inés group (Fig. 6). Marine waters again invaded the geosyncline from the east and eventually spread southward far beyond the limits of Merecure deposition.

In the eastern oceanward portion of the Santa Inés sea and extending westward in a belt along the axis of the geosyncline fine-grained sediments were deposited in moderately deep water and gave rise to the foraminiferal shale of the Carapita formation. In general, west and north of the shale facies and nearer shore, the shallow marine and brackish-water facies typified by the Capaya, Capiricual,

Uchirito, Orégano, and Quiamare formations were formed. At the same time south of the axis of the basin streams from the Guayana shield and from Andean uplift areas far to the west brought in sediments which were laid down on broad coastal flats to constitute the channel and bar sands, carbonaceous marine to brackish-water shales, lignites, and claystones of the Oficina formation. The Suata, Chaguaramas, Santa Lucía, and Cucharo formations represent still more brackish conditions, farther west, and the Bruzual clays represent a fresh-water deposit which may actually have been formed *north* of the marine basin. Along the northern border of the geosyncline and closely adjacent to the mountains thick coarsely clastic facies such as the Caño Dulce beds, and El Pilar, Guanape, and Peña Mota conglomerates were developed at various times.

After attaining a maximum westward extent in Late Oligocene time the Santa Inés sea withdrew somewhat to the east during the early Miocene. In the Middle Miocene there appears to have been further crustal movement. This was in general mild except for local areas along the northern edge of the basin where pronounced uplift followed by deep erosion took place. Within the more central part of the basin deposition was essentially continuous.

A renewed westward extension of the sea in Late Middle Miocene time constituted the Freites transgression which brought moderately deep marine waters at least as far west as the State of Guárico. The axis of Freites deposition lay considerably south of the present area of the Serranía del Interior. Throughout the axial and southern portions of the basin the typical Freites green shales were deposited, but to the north these sediments graded into the gray foraminiferal shales of the uppermost Carapita formation or the shallow marine to brackish-water deposits of the Orégano formation. To the northwest they graded into the less marine upper Quiamare formation sediments. To the northeast peculiar and as yet poorly understood conditions of circulation, turbidity, or chemical composition began to be developed during Freites time and gave rise to the initial deposits of the La Pica formation with its distinctive foraminiferal fauna. Throughout most of the interior of the basin the deposits of Freites time are conformable on those of the underlying formations of the Santa Inés group. However, on local structural uplifts and near the borders of the basin pronounced angular and erosional unconformities are suggested.

Near the close of the Middle Miocene further crustal movement along the northern edge of the basin with a recession of the sea southward was followed by a transgression which carried La Pica facies deposits northward to rest unconformably on Carapita shale at the northern edge of the basin.

During Late Miocene time extensive changes in conditions of deposition took place. The axis of the basin was shifted southward to central Anzoátegui and Monagas, and there was a general replacement of the Freites and La Pica sea environments by brackish to fresh waters forming the inland end of a gulf opening eastward to the Atlantic Ocean. Throughout much of the central portion of the gulf area this was accomplished with no sharp break marking the close of Freites or La Pica sedimentation—only a gradual transition to increasingly sandy and less marine sediments constituting the Las Piedras formation of the overlying Sacacual group. Thus in this part of the basin there is probably local lateral transition not only between Freites and La Pica formations but also between them and the Las Piedras formation of the Sacacual group. However, farther north and west in a broad belt of emergence Santa Inés formations were deeply eroded before the fresh-water deposits of the Sacacual group were laid down upon them with angular unconformity.

Accompanying the shifting of the basin and the general retreat of the sea, there was pronounced late Miocene to Early Pliocene folding and uplift. Much of the major folding and faulting seen in the foothills of the Serranía del Interior and in adjacent areas appears to have been associated with this orogenic period at the close of Santa Inés time. The Anaco trend of folding and faulting in central Anzoátegui, while perhaps initiated in pre-Freites time, owes most of its development to this orogeny. Even the tension faulting of the Greater Oficina area on the south limb of the present structural basin appears in large part to belong to this period of movement.

Sacacual deposition took place largely in a brackish- to fresh-water gulf in central Anzoá-

tegui, Monagas, and Delta Amacuro (Las Piedras formation) and under fresh-water conditions in lakes and rivers near its margins. Fresh-water deposition was particularly prominent along the northern side of the basin where the thick Quiriquire formation was built up from detritus derived from the mountains immediately to the north. The southward-flowing streams from these mountains brought principally material from the Cretaceous and younger formations, but at the same time heavy-mineral data suggest that the rivers must have reached back into the metamorphic areas north of the Serranía del Interior.

In Middle or Late Pliocene time the whole Eastern Venezuela region was uplifted and tilted to the east, and the waters of the gulf were driven eastward to the present delta area. Orogenic movements occurred along the northern border of the basin, and in this area Sacacual beds were folded and eroded before the deposition of the overlying Mesa formation. The northern mountains were uplifted, and the Pleistocene Mesa formation was largely formed by the debris dumped into the basin from these ranges and those of the Andes farther west. A portion of the southern Mesa formation was at the same time derived from mild erosion of the rocks of the Guayana shield.

The mesa deposits are of diverse origins but in large part represent coalescing alluvial fans which built up a vast depositional plain covering the interior of Venezuela and Eastern Colombia, rising high up into the bordering mountain ranges and beyond the present course of the Orinoco River to the foot of the highlands of the Guayana shield. The thickness and coarseness of deposits decreased away from the mountain areas, and in Eastern Venezuela the plain had a very gentle inclination southward and eastward. Probably there was no one master stream through the region at this time corresponding to the Orinoco River of today. Arid to semiarid conditions prevailed, and the waters of the mountain streams were soon lost in their own detritus though in flood times local playa lakes were formed. Eastward drainage in the basin apparently centered at about the latitude of the present Guanipa River.

The history of the existing physiography of Eastern Venezuela may be considered as beginning at about the close of the main Mesa deposition in Pleistocene time. At this time the rate of erosion of the bordering mountains diminished, and an approach to equilibrium conditions was established throughout the basin. Wind erosion was active in this arid region and helped to smooth the surface of the broad interior plains. The same eolian agency produced a concentration of lag and residual gravels which were ferruginously cemented at or near the surface by evaporating ground waters to form the hard ironstone capping of the primary mesa.

At about the close of Pleistocene time further mild uplift of the Caribbean and Andean mountains probably took place and was accompanied by gentle eastward tilting of the basin, perhaps *paripassu* with geosynclinal sinking of the easternmost seaward end of the basin. Climatic changes occurred and with increased rainfall mountain streams began to extend southward into the interior basin where eastward drainage began to develop in conformity with the tilt of the mesa surface and the eastward outlet to the sea. At about this same time a major eastward-flowing stream course may have begun to develop near the northern border of the Guayana shield (ancestral Orinoco). The streams cut through the mesa cap and reworked the Mesa formation sediments on either side forming gentle slopes leading down into the valleys from either side and leaving the original mesa caps only near the interstream divides. During additional periods of aridity and reduced erosion the surfaces of these slopes with their residual gravels were in turn ferruginously cemented by evaporating ground water producing the secondary mesa caps of the present day which tend to slope toward the major streams courses.

Still more recently there was further uplift and tilting of the whole Eastern Venezuela Basin, which together with increased precipitation rejuvenated the mesa streams and initiated the present erosion cycle. The secondary mesa caps were eroded back from the present stream courses and their tributaries to form the characteristic *farallones* bordering these streams today while only remnants of the primary mesas remain (such as the high mesa of the Oritupano area and the hills of San Tomé).

(The presence of many different mesa levels is probably due to local differences in groundwater tables and consequent horizons of cementation.) Even today third-order mesas are locally being formed by mesa outwash within the *farallón* walls of the secondary mesas as a result of most recent rejuvenation of mesa streams.

The increased rainfall of the Recent epoch made mighty streams out of the Orinoco River and its major tributaries. Crustal warpings may even have helped it to capture drainage in the Ventuari-Casiquiare region which formerly was going into the Amazon. The large amount of sediment carried eastward by this system is attested by the size of the Orinoco delta area which has developed at the eastern end of the basin in the shelter of the island of Trinidad in spite of the subsidence of this area indicated by recent sediment 500–1000 feet thick encountered in wells drilled here.

Along the northern border of the basin adjacent to the Serranía del Interior the rejuvenated streams flowing southward out of the mountains cut through the thick mesa sediments and established courses independent of the pre-mesa topography. Thus were developed courses such as those of the Querecual and similar streams which cut across the high ridges of resistant steeply dipping older rocks to get out of the mountains and foothills instead of taking the easier outlets along the strike of nonresistant beds.

While the early drainage of the Eastern Venezuela Basin was probably almost entirely to the east and south a weak spot in the northern mountain barrier was soon developed west of Barcelona where a northward bending of the Serranía del Interior had left only a relatively low barrier of Miocene ridges along the Caribbean coast. At two points, probably localized by transverse faulting, drainage was developed northward across these ridges to the sea. The two resulting northward-flowing river systems (Unare and Aragua-Neverí) have eaten far back to the southward and have captured much of the drainage which formerly had its outlet to the south or east. They have removed the Mesa formation from the broad llanos area of northern Anzoátegui and northeastern Guárico and are actively pushing the mesa front eastward all along the line from Pariaguán to Piripire.

## Structural Geology

The fascinating subject of the tectonics and structural geology of the Eastern Venezuela Basin has been touched on only incidentally and only in connection with its immediate bearing on the stratigraphy and stratigraphic history of the region. A detailed review and study of the information available on this phase of Eastern Venezuelan geology would unduly extend the length of the present report and would also require the introduction of confidential information of direct economic significance which the writer would not be justfiied in presenting for publication. Moreover, a considerable amount of the factual data on structure of this region has already been made available to Bucher and is thus covered in the text of his report and shown on his geologic-tectonic map.

## Economic Geology

The major economic geologic product of Eastern Venezuela is of course petroleum. A discussion of the oil fields of this region is beyond the scope of the present report. Many of the major fields have been described in recent papers published in the Bulletin of the American Association of Petroleum Geologists (Hedberg *et al.*, 1947; Funkhouser *et al.*, 1948). Other mineral resources of Eastern Venezuela which are either being exploited or are at least available in some quantity for exploitation include coal, asphalt, limestone and cement, clay products, building stone, gypsum, gravel, and, road material, sulphur, and salt. Iron-ore deposits at the southern edge of the Eastern Venezuela Basin, in the State of Bolivar just south of the Orinoco River, have recently come into great prominence.

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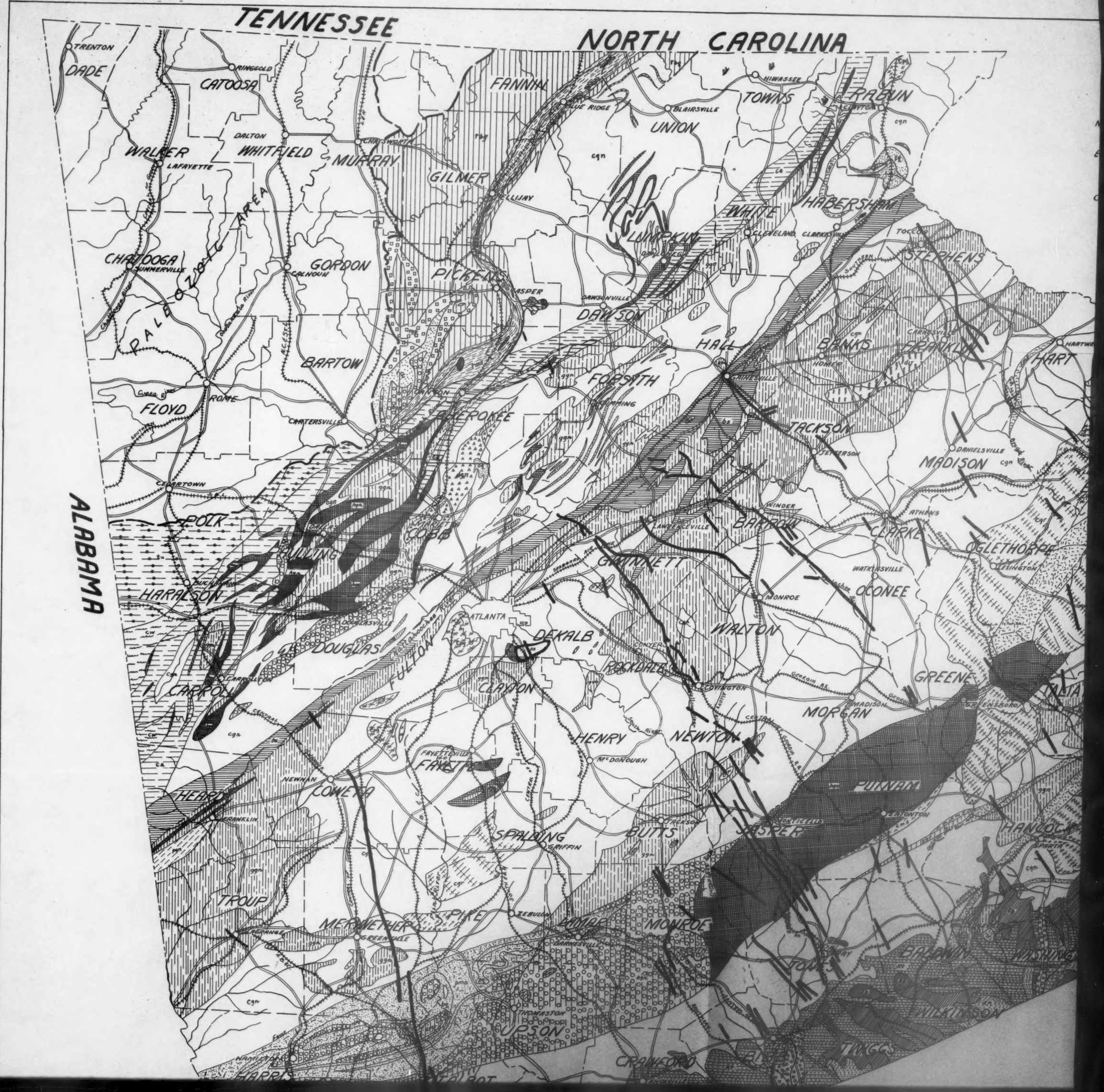
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GWINNETT
LAWRENCEVILLE
BARROW
WINDER
CLARKE
ATHENS
OGLETHORPE
LEXINGTON
WATKINSVILLE
OCONEE
MONROE
HARALSON
BUCHANAN
DOUGLAS
DOUGLASVILLE
ATLANTA
DEKALB
FULTON
WALTON
CONYER
ROCKDALE
COVINGTON
GREENE
CARROLL
MORGAN
MADISON
GREENSBORO
HENRY
NEWTON
McDONOUGH
FAYETTEVILLE
FAYETTE
HEARD
NEWNAN
COWETA
PUTNAM
EATONTON
JASPER
SPALDING
GRIFFIN
BUTTS
TROUP
PIKE
ZEBULON
MERIWETHER
LAMAR
MONROE
GREENVILLE
BARNESVILLE
UPSON
THOMASTON
CRAWFORD
TWIGGS
WILKINSON
HARRIS
TALBOT
MADISON
DANIELSVILLE
Chattooga Fault
Chattahoochee River
Coosa River
Ocoee River
Etowah River
Ogeechee River
Savannah River
Southern Ry.
Central of Ga. Ry.
Georgia RR.
Seaboard A.L.
Atlanta & West Point
Western & Atlantic
Louisville & Nashville
Georgia Ry.
Southern Ry.
Blue Ridge
Tocca
Carnesville
Homer
Chinesville
Lawrenceville
Winder
Athens
Lexington
Watkinsville
Monroe
Madison
Greensboro
Eatonton
Monticello
Griffin
Fayetteville
McDonough
Newnan
Carrollton
Douglasville
Buchanan
Cedartown
Rome
Summerville
Trenton
Ringgold
Dalton
Chatsworth
Ellijay
Blairsville
Clayton
Clarkesville
Cleveland
Dawsonville
Jasper
Canton
Jackson
Barnesville
Zebulon
Thomaston
Talbotton
Greenville
LaGrange
Franklin
Carrollton
Conyers
Covington
Madison
Sparta
Hartwell
Cgn
Ry
Ls
Ss
Sh
Ls
Ss
Sh
Ls
Ss
Sh

NORTH CAROLINA

SOUTH CAROLINA

TENNESSEE

FANNIN · TOWNS · RABUN · UNION · MURRAY · GILMER · WHITE · HABERSHAM · LUMPKIN · STEPHENS · PICKENS · DAWSON · HALL · BANKS · FRANKLIN · HART · FORSYTH · CHEROKEE · JACKSON · MADISON · ELBERT · CAMPBELL · COBB · GWINNETT · BARROW · OGLETHORPE · CLARKE · OCONEE · WILKES · LINCOLN · DOUGLAS · FULTON · DEKALB · ROCKDALE · WALTON · GREENE · TALIAFERRO · McDUFFIE · COLUMBIA · COWETA · FAYETTE · HENRY · NEWTON · PUTNAM · HANCOCK · WARREN · GLASCOCK · JEFFERSON · RICHMOND · SPALDING · BUTTS · JASPER · MERIWETHER · PIKE · LAMAR · MONROE · JONES · BALDWIN · WASHINGTON · UPSON · WILKINSON · TWIGGS

BLUE RIDGE · BLAIRSVILLE · HIWASSEE · CLAYTON · CHATSWORTH · ELLIJAY · CLEVELAND · CLARKESVILLE · TOCCOA · JASPER · DAWSONVILLE · CANTON · CUMMING · GAINESVILLE · HOMER · CARNESVILLE · HARTWELL · DANIELSVILLE · ELBERTON · LAWRENCEVILLE · WINDER · ATHENS · LEXINGTON · WATKINSVILLE · MONROE · ATLANTA · DOUGLASVILLE · CONYERS · COVINGTON · MADISON · GREENSBORO · CRAWFORDVILLE · WASHINGTON · LINCOLNTON · FAYETTEVILLE · McDONOUGH · GRIFFIN · ZEBULON · GREENVILLE · BARNESVILLE · THOMASTON · MONTICELLO · EATONTON · SPARTA · WARRENTON · LOUISVILLE · AUGUSTA · SAVANNAH

## LEGEND

### COASTAL PLAIN

- RECENT
- MIOCENE — HAWTHORNE
- EOCENE — BARNWELL
- EOCENE — McBEAN
- CRETACEOUS — CUSSETA
- CRETACEOUS — TUSCALOOSA

### CRYSTALLINE AREA

- LITTLE RIVER
- TALLADEGA
- MURPHY MARBLE
- MICA SCHIST
- GRAPHITIC PHYLLITE
- BIOTITE GNEISS
- QUARTZITE & CONGL.
- BREVARD SCHIST
- MARBLE IN BREVARD
- MANCHESTER & SPARKS SCHIST
- QUARTZITE
- MUSCOVITE SCHIST
- CANTON SCHIST
- WEDOWEE
- ASHLAND SCHIST
- QUARTZITE
- BIOTITE GNEISS & SCHIST
- BIOTITE & MUSCOVITE GRANITE
- PORPHYRITIC GRANITE
- PERIDOTITE, SERPENTINE
- PORPHYRITIC GRANITE GNEISS
- GRANITE GNEISS
- DIORITE INJECTION GNEISS
- GARNET-PYROXENE GRANITE
- AUGEN GNEISS
- HORNBLENDE GNEISS
- DIORITE GNEISS
- DIABASE

**GEOLOGIC MAP OF AREA**

Showing the occurrences of diabase. (Geology taken from Georgia geologic map prepared by Division of Mines, Mining, and Geology - 1939)

SCALE

5 0 5 10 15 20 25

**GEOLOGIC MAP OF AREA**

Showing the occurrences of diabase. (Geology taken from Georgia geologic map prepared by Division of Mines, Mining, and Geology - 1939)

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# DIABASE OF THE GEORGIA PIEDMONT

BY J. G. LESTER AND A. T. ALLEN

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## ABSTRACT

The crystalline area of Georgia, comprising more than 17,000 square miles, was studied for the occurrence of diabase. Fifty-seven new dikes were mapped and specimens from them examined under the microscope. No sill-like bodies were encountered and no flows or remnants of flows exist in the area.

No definite age can be given to the diabase in Georgia, but on the basis of contacts with certain rock units and deep well samples from Florida, the age has been set as late Triassic.

Contact metamorphism is seldom present, and then only slightly so; no mineral concentrations were found; and the dikes have little, if any, influence on the drainage patterns of the region.

## INTRODUCTION

The diabase dikes in Georgia have been studied previously by McCallie (1901), Watson (1902, p. 81, 96), and Hewitt and Crickmay (1937). The writers present this report as a more detailed study than previous ones on the occurrences of diabase in Georgia.

The area of this study (Fig. 1) includes 55 counties and comprises more than 17,000 square miles, slightly less than a third of the total area of the State of Georgia. It includes that part of Georgia bounded by the Coastal Plain on the south, the Alabama State line on the west, the Paleozoic front on the northwest, the Blue Ridge highlands on the north and northeast, and the Savannah river on the east. Fifty-seven new dikes with a total outcrop length of more than 280 miles have been mapped and studied. Fifty-five previously known dikes with a total outcrop length of 256 miles, indicated on the Georgia Geologic Map of 1939, have also been studied. During the course of the field work from 1947 to 1949, 173 separate rock samples were taken from which almost 400 thin sections were prepared.

The dikes were located by traversing all roads normal to the regional strike and by stereoscopic examination of aerial photographs. The continuity of the dikes was determined by traversing along the strike between outcrop points.

## ACKNOWLEDGMENTS

Thanks are extended to A. S. Furcron of the Georgia Division of Mines, Mining, and Geology for memoranda concerning dike occurrences; to the U. S. Corps of Engineers for aerial photographs; to Mr. Herman Gunther of the Florida Geological Survey for deep well samples.

The project was made possible through a grant-in-aid allocated by the Research Com-

FIGURE 1. INDEX MAP Shading shows area investigated

mittee of Emory University and from funds made available jointly by the Carnegie Foundation and that University. The authors and not the University or the Foundation are solely responsible for all statements made in this report.

## PHYSIOGRAPHY

The map area includes the major part of what has been called the Crystalline Belt of Georgia or the Central Upland by LaForge (1925) and is an ancient peneplane of moderate relief characterized by gently rolling hills and fairly broad stream divides. The general slope of the area is to the south, southeast, and southwest. The drainage patterns are dendritic to subdendritic and all streams seem to be either consequent or superposed.

The general level of the peneplane is occasionally interrupted by solitary hills locally called mountains with altitudes ranging from

900 feet above sea level to 1809 feet above sea level.

The average annual rainfall exceeds 60 inches and the climate is mild for most of the year. As a result of climatic conditions, a heavy residual soil mantle, varying from a few feet to more than 200 feet in thickness, has developed.

The dikes do not materially affect the topography, although the course of the larger ones can be traced as low, sinuous ridges. The surface of the ridges is covered with countless rounded boulders of varying size resulting from the spheroidal weathering of the diabase. On the broad flat divides between the major streams, the dikes are not as prominently exposed and can be traced only with considerable difficulty. In some instances the only evidence which can be relied upon is the occurrence of randomly scattered boulders turned up in plowing, and the changes in the color of the residual soil.

The rocks of the area are meta-sediments and meta-igneous rocks with extensive outcrops of granites, numerous diabase dikes, and a few other basic dike rocks.

Many of the larger masses of igneous and meta-igneous rock stand above the general level of the region as monadnocks; the most prominent are Stone Mountain, Kennesaw Mountain, Little Stone Mountain, and Panola Mountain. Smaller monadnocks such as Alcovy Mountain, Cornish Mountain, and Graves Mountain are composed essentially of complexes of igneous and metamorphic rocks.

Age relationships of the numerous formations in the crystalline area have not been accurately determined, and the twelve formations intruded by the diabase are shown on Plate 1.

## Diabase Dikes

### *General Statement*

In general, the dikes strike NW-SE but the strikes tend to swing toward the north in the western part of the area and toward the west in the eastern part. Only the shorter dikes have a constant strike; the longer ones are sinuous and the strike changes along their course. Dips vary from 75° to vertical, the majority being vertical. All exposures of the diabase occur as dikes and no sills or sheets are encountered.

### *Petrography*

Fresh surfaces of the rock show it to be fine-grained and dark gray to black. The black color of the fresh rock changes to a dark brown upon slight weathering and to a yellowish-brown upon more advanced weathering.

Microscopically the rock is seen to be composed essentially of plagioclase feldspar and pyroxene with smaller and irregular amounts of olivine, hornblende, magnetite, and pyrite. Secondary minerals include uralite, kaolinite, and iddingsite.

The plagioclase ranges from $ab_{20}an_{80}$ to $ab_{30}an_{60}$ with $ab_{30}an_{70}$ being the dominant feldspar. The plagioclase occurs as microlites which vary in a single section from a fraction of a millimeter to 4 millimeters long and from a tenth millimeter to 1½ millimeters wide. They are frequently seen in a splintery and branching condition and may exhibit rough alignment as a result of flow. Clustering is common and rosettes resulting from clustering around a center are frequently seen. Both ophitic and diabasic structure occur.

The pyroxene is predominantly a pale-green to colorless augite which is non-pleochroic and varies from subhedral to anhedral. The profusion of microlites frequently destroys the crystal outline to such an extent that the pyroxene assumes the appearance of filling the interstices between the microlites. Hypersthene is occasionally seen and is readily distinguished from other pyroxenes by its lower birefringence and the close association of magnetite along cleavage and fracture planes. The pyroxenes compose from a half to two-third of the coarser-grained varieties. An occasional crystal of enstatite is seen.

Olivine is not encountered in all sections and seemingly does not occur in all dikes. Sections from the dikes of the central zone show olivine more frequently than sections from other areas. When seen it exhibits euhedral to subhedral form, high relief, high birefringence, and irregular fractures. Occasional feathery reaction rims of a reddish-brown iddingsite are seen.

Hornblende is not common but is encountered in some sections where it assumes euhedral to subhedral outline, and exhibits characteristic cleavage and weak pleochroism. It frequently is intimately associated with augite, from which it seems to have been derived. Uralite, also an

alteration product, is occasionally seen as light-green patches of anhedral form scattered through the field of the section.

Magnetite in subhedral to anhedral grains, probably after the pyroxene, is frequently plentiful in certain parts of the section and under uncrossed nicols give the slide a mottled appearance. In other sections it is primary in origin and occurs as euhedral to subhedral crystals.

Kaolinization is frequently observed in the sections of coarse textured rocks and is so intense that areas of the section are obscured to the extent that determinations are difficult or impossible.

Biotite is uncommon but when seen is light brown, slightly pleochroic from light to dark brown, and in minute shreds.

Textures range from very coarse to very fine and some of the aphanitic varieties show trachitic structure. Porphyritic texture is occasionally encountered in which the phenocrysts are pyroxenes and olivine. A small number of sections exhibit a sub-mosaic texture in which the feldspar laths are nearly equidimensional and the pyroxenes appear as anhedral grains interspersed at random among the laths.

The diabase of Georgia is remarkably free of cavities. Only one dike, in Lincoln County, shows spherulitic cavities which have been filled, for the most part, with hydrothermal silica and calcite. No calcite derived from alteration of pyroxenes, as described by Goodspeed (1934) in the diabase of Washington state, was encountered.

As a whole, the Georgia diabase is monotonously uniform in composition, appearance, structure, and alteration.

### *General Description of Occurrences*

The main dikes have a constant dip toward the east of 75° to 90° from the horizontal. Some of the smaller dikes and some of the satellite branches occasionally assume dips as low as 40° which tend to steepen slightly as the minor dikes join the major dike.

The prevailing strike is toward the northeast in the western part of the area and toward the northwest in the central and eastern parts. A few trend in an east-west direction, some in a north-south direction, and some of the more persistent dikes change their trends from northeast to northwest with gentle curves.

The dikes crop out from a few hundred feet in length to as much as 35 miles. The longer dikes are unique because most of them are discontinuous. Widths range from a few inches to several hundred feet and may vary within a single dike because of the tendency of the dikes to pinch and swell along the length.

Discontinuity occurs for the following reasons: The dikes may pinch out completely and then expand again with the same strike and same structural characteristics. They may be broken by faulting. They may rise and fall in a vertical sinuosity so that no trace of it can be seen for short distances. They may be blocked out by another rock formation too massive and too resistant for the diabase to penetrate. Transection streams may bury the dike beneath alluvium. Differential weathering may affect gaps in the dike and the resulting residual soil mantle completely masks the fresh rock. All conditions of discontinuity are seen in the area studied.

A few dikes send off branches but in only one case observed did the branches exceed two in number. A large dike in Barrow County has twelve smaller dikelets on one side of it, and their dips, proximity, texture, and composition indicate that they join the large dike below the exposed part. The main dike and its off-shoots form a persistent ridge of low relief, covered with sparse vegetation, which can be traced for more than 5 miles.

The number of dikes shown on the map of the area does not include the branches and off-shoots which are too small to show on the map, nor those dikes completely buried beneath the thick residuum. In several instances, small dikes can be seen in the lower parts of highway cuts when no single indication of a dike is visible on the surface. Careful observation will reveal their presence as a sudden change in soil color which is often masked by vegetation.

Jointing is always characteristic of the rock, but no single type of jointing predominates. Some dikes exhibit joint blocks of definite rhombiform shape, others show distorted wedge-shaped blocks, and one dike near Gray, Georgia, in Jones County exhibits octahedral-like joint blocks. The major joints trend at angles of from 75° to 90° from the contact of

the dikes and the country rock. The size of individual joint blocks varies from an inch or two on an edge to blocks several feet on an edge and is proportional to the size of the dike and the rapidity of cooling. When blocks are found in place in the dike, they normally exhibit nearly perfect jointing.

Dikes wider than 10 feet invariably show a chill zone along the borders which rarely exceeds an inch or two in thickness. Some smaller dikes are so fine-grained that chilling is indicated through the entire width. Jointing and other structural features are seemingly unaffected by the chill zone.

Some of the formations intruded by the diabase show a change in color and in some instances a very slight baking effect similar to that described by Noble (1914) in the Grand Canyon shales.

Some significant facts and important inferences can be drawn from the study of the intrusion of the diabase into two dissimilar rocks: the Stone Mountain granite and the Lithonia granite gneiss. In the exposure of the Lithonia granite gneiss at Camak, Warren County, Georgia, the diabase transects unaltered gneiss with seemingly little effect other than a color change which is seen as a band only a fraction of an inch wide. In the Stone Mountain area, the diabase cuts a weathered portion of the granite which exhibits a narrow border of color change plus a thin zone of baking an inch or so wide. Although the contact phenomena are minor, it is pertinent that a zone of baking occurs. These observations seem to indicate that the Stone Mountain granite may have been partially weathered prior to intrusion of the diabase and the clayey material of the partially weathered rock was baked; also that the subdued topography of the Crystalline area may have begun to form before the diabase intrusion; and finally, that the intrusive force and the contained gases and solutions of the diabase had about spent themselves in that they were able to penetrate a weathered, porous zone of granite and were unable to penetrate the unweathered, finer-grained gneiss.

In a few instances, fragments of the country rock have been trapped within the diabase and it is thought that the force of the intrusion of the dikes showing such inclusions was relatively high. However, no such forces as indicated by Ransome (1903) in the diabase of the Globe Copper District of Arizona are thought to have been in action in these dikes.

### *Weathering and Geomorphic Relations*

The processes of weathering advance rapidly under the climatic conditions existing in the Piedmont of Georgia at the present time. A study of the weathering reveals that the tendency for exfoliation to take place, rounded or rhombiform boulders to develop, or for the diabase to completely disintegrate is directly related to the topography and to the amount and velocity of runoff.

On the broad, high stream divides, even the larger, more massive dikes are traced with difficulty. The fact that the dikes can only be seen here as an occasional boulder or as a distinctive change in soil color can be explained by the fact that the residual soil in these areas accumulated to a marked degree. The presence of the thick soil retains moisture and causes the diabase to be constantly saturated and subjected to chemical changes. The pyroxenes, olivine, plagioclase, and hornblende are oxidized by the downward percolating ground water and there is developed a thick clayey soil which is high in the iron oxides.

In road cuts which cross the dikes in the higher areas, these phenomena can be observed. Joints can be seen cutting the diabase into blocks of various sizes. Weathering advances on all the blocks from all surfaces intersected by the joints, and the degree of weathering decreases from the outer to the inner part of the dike. It is frequently noticed that little or no exfoliation is evidenced by dikes subjected to this type of weathering.

In areas where the gradient is steeper and runoff removes the residual soil as fast as it is formed, rhombiform or spheroidal boulders can be expected. These shapes are characteristic of the dikes when the outcrops are afforded the opportunity of drying between periods of wetting. Under these circumstances, the residual soil is not present in sufficient amount to retain much moisture and the blocks tend to become rounded by the freer percolation of water. The corners of the angular blocks formed by the intersecting joint planes are the first to be destroyed. The clay is removed as fast as it is

formed, and the repeated wetting and drying causes the blocks to progress from angular forms to spheroidal forms.

Exfoliation of the diabase seems to be both a chemical and a physical process. Apparently these processes take place more readily on the coarser varieties of diabase. In dikes in Jones and Bibb counties, weathering shells have developed until as many as 25 separate shells can be counted encircling a single block; the unweathered central mass often being no larger than an inch in diameter. The spheroidal boulders thus developed can be observed long distances from their source. Watson (1898) described similar weathering in the diabase of Virginia.

Even though the rock is exceedingly tough and resistant to fracture when fresh, its rapid weathering in Georgia is so characteristic that it is shunned for all types of permanent construction work and is used only locally in rock gardens, property line walls, and foundations for minor farm buildings.

### *Distribution*

The dikes are not evenly distributed over the region; some parts show heavy concentration of dikes, others show them loosely scattered and still other parts are seemingly free of them. There are, broadly speaking, five zones of dike concentration which strike in general from N 10° W on the west to N 35° W on the east. The zone showing the greatest number of dikes occurs near the central part of the region and forms a general area of divide between the Ocmulgee and the Oconee drainage basins.

The areas between the dike zones, seemingly barren of diabase, can be roughly correlated with the mica-producing pegmatite areas of Georgia as described by Furcon and Teague (1943).

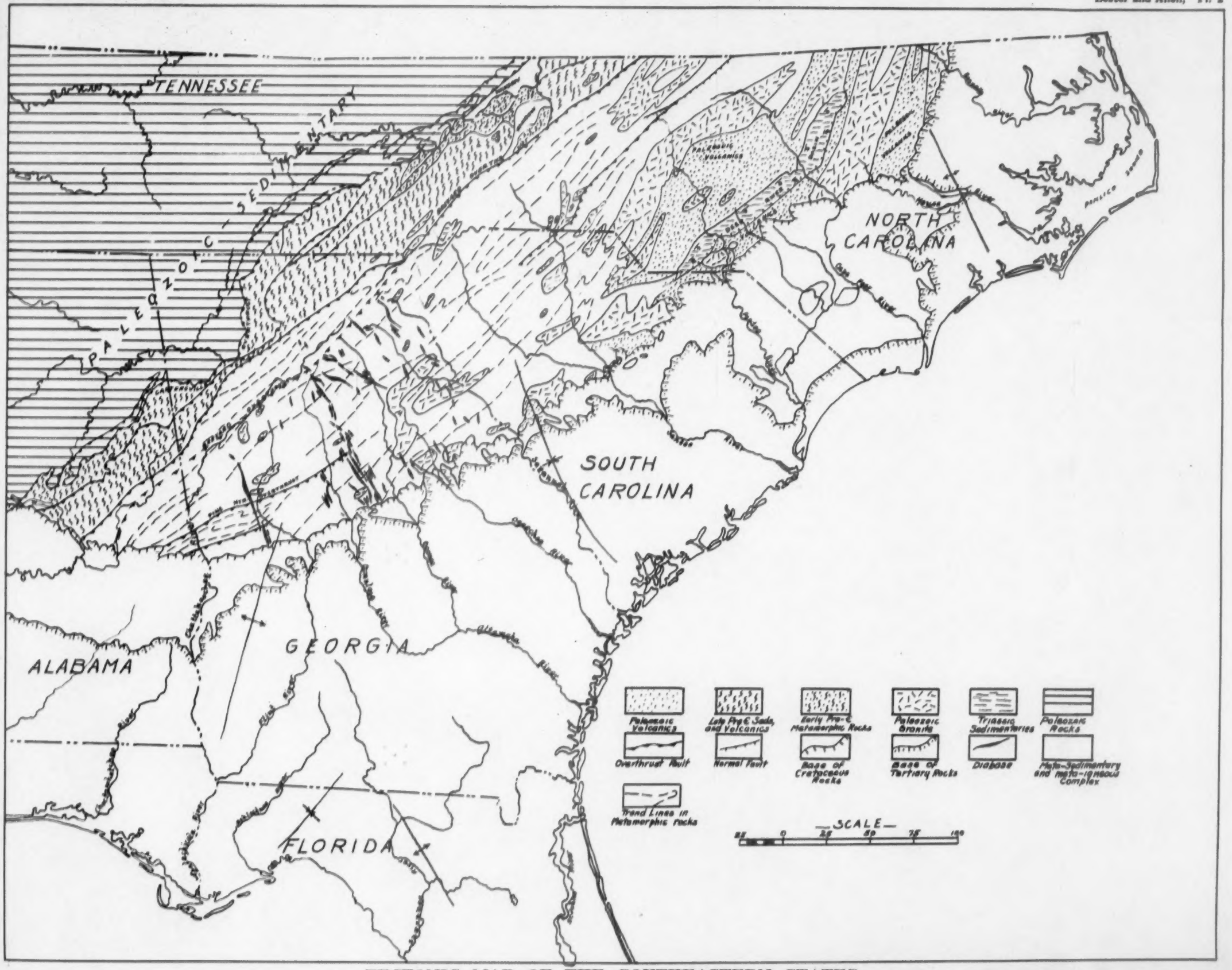
The dikes range in thickness from an inch or so to several hundred feet and in length of outcrop from 1000 feet to 35 miles. The larger dikes exhibit pinching and swelling and wide variations in strike but little if any change in dip. The most remarkable feature of the long dikes is the relatively narrow width prevailing throughout most of the length.

### Structural Relations

Examination of the tectonic map (Pl. 2) will reveal the following facts: (1) The dikes do not extend beyond the Brevard Overthrust; although in five instances dikes extend into the formation shown on the geologic map as Brevard schist, they do not extend beyond the line of overthrusting. (2) In general, the major trend of all dikes is almost normal to the trend of the Brevard Overthrust. (3) The dike zones are essentially parallel to the major anticlinal folds of the region. (4) The zones in which numerous dikes occur are composed of relatively short dikes for the most part and are the zones in which the greatest variation of strike occurs. (5) The number of dikes diminishes from the median line of the area to the east and to the west. (6) If dike occurrences are examined in states other than Georgia, through which the Brevard Overthrust passes, it is believed that the strike will, in general, conform to the attitude found in the Georgia area. (7) Dikes near the Fall Line trend normal to it.

One gains the impression that some structural relation exists between the Brevard Overthrust and the younger diabase dikes. Possibly this zone acted as a barrier beyond which dikes do not extend, the only exceptions being a dike in Cherokee county and one in Rabun county. However, recent work by Grant (1949) shows that no single thrust fault exists but rather that a belt of numerous small faults comprise the so-called Brevard Overthrust in Georgia. If such is the case the rock may have been so thoroughly metamorphosed prior to the intrusion of the diabase that the dikes were not able to penetrate it with the intrusive forces they possessed within themselves. Since all dikes are filled fissures, it is logical to assume that no fissures extended across the Brevard Overthrust zone.

In the Amplett Mica Mine near Ball Ground, Cherokee County, and in Jasper County, diabase intrudes granitic pegmatite. These are the only localities in the area where age relations between diabase dikes and pegmatites have been found. If these can be taken as criteria for all pegmatites, they are definitely older than the diabase intrusions—possibly as old as late paleozoic and associated with the Appalachian revolution. Obviously the folding along the




**TECTONIC MAP OF THE SOUTHEASTERN STATES**

Showing major geologic divisions and structure of crystalline belt. (Structure from Tectonic Map of the United States, American Association of Petroleum Geologists)

Fall Line and on the Coastal Plain during Eocene and Miocene time is much younger. In the light of these relations, it would seem evident that the crystalline area and the area overlain by the Cretaceous and younger sediments has been one of unrest for a long period of time.

Since the strike of the diabase dikes is normal to the Fall Line and to the hinge line of pre-Cretaceous downwarping, possibly the fractures occupied by the diabase are associated with such movement. On the theory that downwarping causes tensional cracks to be formed parallel to the axis of downwarping, which in turn are filled with basic igneous material in order to maintain isostatic adjustment, it is postulated that the areas of dike concentration were the axes of northwest-southeast trending synclines which existed simultaneously at the time of the diabase intrusion. A force couple, acting along the southern edge of the Piedmont, could develop folds which would plunge to the northwest and have longitudinal axes almost normal to the Fall Line. This condition would be similar to that obtained by compression of a piece of cloth cut on a bias which would essentially parallel the trend of the Fall Line. In the light of this explanation, the Brevard Overthrust assumes a minor position and the Fall Line assumes a major role in the arrangement, position, and extent of the dikes.

The authors believe that the source of the diabase was a relatively deep magmatic basin. The flexural forces acting on the Appalachian region opened fissures along which the diabase could pass freely and easily. Thom (1937) explains the gravitational "highs" which diverge westward from the Fall Line extended southward as intrusive bodies. Tectonic forces, perhaps of greater intensity in the New England region, caused the great extrusive sheets of the northeastern part of the United States. The same forces generated a series of folds in the Piedmont whose axes trend northwest-southeast and at the same time brought about a migration of the diabase magma from the northeast toward the southeast which was intruded as dikes in the fissures opened by the folding. No sheets, remnants of sheets, flows, nor sills have been found in the Georgia area.

## Age Relationships of the Diabase

The diabase of the eastern United States has been placed in the Triassic by Darton (1880), Davis (1898), Phillips (1899), Leonard (1901), Roberts (1928), Smith (1931), and others.

All of these workers, reporting on areas from Connecticut to Virginia, were able to place its age as a result of stratigraphic position and relationship. In the Georgia area where the rock units are all metamorphosed sedimentary or igneous rocks, the age of the diabase which intrudes them can only be established by indirect evidence.

Of the twelve rock units or rock groups intruded by the diabase, eleven have been dated as pre-Cambrian; Stone Mountain granite and the granite pegmatites have been dated as late paleozoic by Watson (1902, p. 279) and Lester (1938).

Conclusive evidence of the age of the diabase has been sought along the Fall Line, which marks the contact between the crystalline rocks of the Piedmont and the sediments of the Coastal Plain. Such evidence is lacking. In fact, no single dike was found which could be traced to a contact with the sediments, and no sediments of the Coastal Plain were found which contained pebbles of diabase. If such sedimentary beds exist, they are so deeply buried beneath the Tuscaloosa formation (early Cretaceous) that exposures of them do not occur.

Recently the writers were supplied with certain well samples taken from deep wells in Florida in which diabase was encountered below the base of the Cretaceous. One such sample was taken at a depth of 11,993 feet and possessed essentially the same minerals as the diabase in the central zone of dikes in the Georgia Piedmont.

## Summary

Fifty-seven diabase dikes were located, mapped, and studied both megascopically and microscopically. The rock was found to be remarkably uniform in composition and remarkably free of contact phenomena. Five zones of dike concentration were delineated, the most prominent zone lying almost in the center of the area. The dikes have little effect

on the major drainage patterns. Although some of the minor tributaries seem to have cut their channels under the influence of the diabase, others are not influenced by the diabase.

It is believed that the dike zones are directly related to downwarps during post Paleozoic time which opened fissures in the crystalline rocks. As far as is known, all dikes originated from the same magmatic source and were intruded at relatively the same time as a result of similar conditions. The age is set as Triassic although no conclusive data can be furnished.

In no case was it found that the diabase exhibited either intraformational or peripheral mineral occurrences other than those normally found within the body of the rock itself. Metallic minerals in appreciable amount and contact minerals are absent.

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# SEA BOTTOM OFF THE COAST OF CALIFORNIA

## A review of Special Paper 31, Geological Society of America, by Shepard and Emery, 1941

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### INTRODUCTION

I propose in this paper to sketch and classify the salient features of the sea bottom off the coast of California, and to discuss some questions as to their genesis. I do not contribute anything new to the observational data concerning the phenomena described. My information as to these is set forth in the recently published monograph by Shepard and Emery (1941), particularly in the four charts which accompany the monograph. The present paper is thus a review of their work. I do, however, offer a classification of the features which appear on the charts, which is different from theirs, and propose another interpretation as to the genesis of the most important feature of the charts. The charts give us for the first time a good picture of the physiography of the sea bottom to the west of California, and owe their excellence, of course, to the vast amount of sounding done by the U. S. Coast and Geodetic Survey at the behest largely of Mr. Shepard. It is a joy to behold a great government bureau, the functions of which are primarily utilitarian, devoting so much of its energies to the solution of a fundamental scientific problem like that of the sea bottom; and it is no less a joy for the geologists of this country that they have so competent an agency as the Geological Society of America ready and willing to undertake the publication of this protracted survey.

The readers of this paper would do well to have before them the four charts mentioned above, and the use of these charts will be greatly enhanced if they are mounted together on one sheet. I hope my readers will be merciful. Although the discussion which follows deals with an important problem, geological evidence is woefully scant, and the conclusions arrived at are, from the author's point of view, little more than formulations of hypotheses.

## General Features of the Sea Bottom

The salient features of the sea bottom displayed by the charts of Special Paper 31 are deployed in parallel zones having an aggregate width of more than 60 miles off the coast, and parallel to it. The first of these, going seaward, is the Continental Shelf extending without a break for the entire length of the state and far beyond both to the north and to the south. In width it varies from less than a mile to about 35 miles. Its upper limit is, for the present position of sea level, practically the shore line. Its lower limit for purposes of mapping is contour −600 feet.

The second zone (II of this text) is the most noteworthy and important feature of the 60+ mile belt. It comprises a strip of sea bottom which is persistently linear and has a relatively steep slope seaward. It is called the Continental Slope by Shepard and Emery. The upper limit of this steep slope is contour −600 feet, which, as stated above, is also the seaward limit of the Continental Shelf.

Although the Steep Slope is remarkably constant in orientation, it is subject to certain interruptions which may be here mentioned: These comprise the gorges which traverse the zone roughly normal to its strike, and some transverse faults which locally offset its course. The most important interruption, at Monterey Bay, may justify the subdivision of this remarkable feature into two structurally distinct parts. The embankments, which are the chief feature of Zone III, may be considered an interruption on the dip, or at least the declivity of the Steep Slope; since the latter emerges everywhere from beneath the upper edge of the embankments. There are, however, embankments of two diverse origins. The prevailing type of embankment has its highest point, or apex, at or in the vicinity of the mouths of the gorges, or 'canyons' which cut through the Steep Slope transversely. They correspond in position and in conical slope to the deltas, or alluvial fans, of the rivers of the land. I shall call them for lack of a better term deltaic embankments. The less common type of embankment encircles the seamounts of which five are recognized by name on the charts. These are probably composed of volcanic ejecta emanating from the seamounts. In the proximity of seamounts it may be difficult to distinguish their encircling embankments from those of the deltaic type.

Zone III of the 60+ mile belt comprises a strip of sea bottom of unknown width extending out from the lower part of the steep slope and mantled by the embankments referred to in the last paragraph. It is the chief site of deposition of the erosional waste from the canyons which traverse the Steep Slope; and the relationship of adjoining deltaic embankments is that of overlap and interdigitation.

In between the Continental Shelf and the present top of the Steep Slope is Zone IV which is in strong contrast with both Zone I to the east of it and with Zone II to the west. It has a seaward declivity much greater than that of the shelf and much less than that of the Steep Slope. The zone is not persistent throughout the length of the coast, and its width is much more variable than that of the two flanking zones. It is entirely lacking from those parts of the 60+ mile belt in which the top of the Steep Slope coincides with the outer edge of the Continental Shelf at the common boundary of contour −600 feet. This intermediate, nonpersistent zone is traversed by numerous canyons, large and small, and its general surface appears to be one of uneven erosion.

## Some Details of the Four Zones

### Zone I—The The Continental Shelf

*General description.*—The features of the Continental Shelf are the most sharply defined and most constant of the four zones. The landward boundary of the shelf is not a contour. In any vertical section it begins at the bottom of the beach slope, or, if there be no beach but only a sea cliff of resistant rock, it begins at the toe of the talus of the cliff. At the mouth of a river entering the sea, the position of the landward edge of the shelf cannot always be distinguished in the field from the surface of a delta which upstream may be situated above sea level.

The seaward limit of the shelf is sharply marked by the difference in the spacing of the contours above and below what is taken to be its outer edge. This contrast places the edge at

contour −600 feet for that portion of the coast between the northern boundary of California and Pt. Sur, a distance of 450 miles. But this condition ends abruptly at Pt. Sur where the limiting contour of the shelf becomes −300 feet, and so continues southerly to the Mexican boundary and beyond. In looking for an explanation of this change of depth of the seaward limit of the Continental Shelf an examination of the chart immediately south of Pt. Sur shows that within a coastal stretch of 20 miles south of that point the branches of three notable 'canyons', named Sur, Partington, and Lucia, have cut back shoreward so that the width of the shelf at contour −300 feet is only about half a mile at the head of the Partington Canyon. It is clear that erosive action at the heads of these canyons has reduced the width of the shelf and the depth of its outer edge. This reduction has exceeded the tendency of the shelf to extend seaward into water deeper than 300 feet. The next canyon southward is Arguello Canyon in the latitude of Pt. Conception. This also has a group of tributaries which have cut back shoreward beyond contour −600 feet but have not yet reached contour −300 feet. This latter contour is, however, serrate in plan for 30 miles on either side of Pt. Conception, and the edge of the shelf is evidently crumbling. From Pt. Hueneme to the Mexican boundary there are nine submarine 'canyons' named on Chart I: Hueneme, Muga, Dame, Santa Monica, Redondo, San Pedro Valley, Newport, La Jolla, and Coronado, most of which make deep embayments in contour −300 feet and head in close to shore. The shelf for this stretch of shore is also in course of destruction. Even the −300 foot contour is not fully maintained as the seaward boundary. It is yielding to the erosional removal by way of the 'canyons' named. If the supply of sediment from the land were more abundant, the story would be different, and the shelf wider. The streams which bring material from the land for the construction of the shelf are nearly all small and intermittent. Under a rainy climate the shelf would probably maintain a greater width. In the width of the shelf we may have an integrated measure of climate. For any geological epoch, such as the *present*, the sea level is constant and climate is the chief variable factor in the building of the shelf.

A very disturbing factor in the use of such a measure, however, is a change in the depth of water off shore by subsidence of the sea bottom. In a later part of this paper it is pointed out that the archipelago off the coast of Southern California is due to a subsidence of the sea bottom which faded out to the north in the vicinity of Pt. Sur. This depression increased the depth of water on the Continental Shelf and promoted its partial destruction by small tributaries of transverse submarine canyons, a process which appears to be still active.

*Thickness of the Shelf embankment.*—Since the Continental Shelf is the top of a marine embankment and is therefore a geological formation, the question of how thick the embankment may be is of interest to geologists. So far as I have discovered, the only partial answer to this question is that the gorge at the Golden Gate has a rock bottom at a depth of at least 382 feet[1] below present sea level. In the vicinity of the Golden Gate, it is probable that the embankment has grown from nothing to a thickness of 382+ feet as the sea level rose with the waning of continental glaciation. But 30 miles due west of the Golden Gate, at the Farallon Islands, the thickness of the embankment is again zero.

*Gradient of the Shelf.*—The slope of the Continental Shelf from the shore to its seaward limit is also a matter of interest to geologists, and it may be opportune to offer a word of comment concerning this feature. In order to get some figures for the measure of the slope normal to the general trend of the coast, I have taken a series of 16 points on the shore line of the charts, half a degree of latitude apart and extending south from the northern boundary of the State to Pt. Conception. From these points I have drawn straight lines across the shelf normal to the coast line and on these normals have measured the distances from the shore to contour −300 feet.

For the 16 normals on which this measurement was made, the respective seaward gradients of the shelf in feet per mile are: 23, 20, 30, 31.5, 46, 86, 54, 60, 43, 18, 50, 240, 86, 100, 37

[1] As determined by the U. S Coast and Geodetic Survey.

66. Of these, I have rejected the gradient of 240 feet per mile as abnormally steep. The mean of the remaining 15 measurements is 50 feet per mile of horizontal distance across the shelf from the shore to contour −300 feet. Measured in the same way and on the same normals to the coast line, the gradients in feet per mile of the shelf between contour −300 feet and contour −600 feet, from Lat. 42° N, to Lat. 37° N, are: 75, 46, 66, 126, 300, 48, 43, 26, 21.4, 50, 100. Here the gradient of 300 feet per mile appears to be abnormally large and it is rejected in getting a mean value for the seaward slope of the shelf. The mean is 60.1 feet per mile for the portion of the slope between contours −300 and −600 feet. It thus appears that this outer portion of the shelf is somewhat steeper than that bordering the shore. The general profile transverse to the shelf is probably slightly convex upward.

Similar measures for the gradient taken as a whole between the shore to contour −600 foot, from Lat. 42° N to Lat. 36° 15′ N, and measured on the same normals to the shore line, are as follows, in feet per mile: 35.3, 28.6, 40, 50, 80, 61.5, 48, 35.3, 28.6, 27.2, 75, 171. The last of these figures is regarded as abnormal and unreliable since the edge of the shelf, where cut by the normal on which the measure is taken, is situated between two tributaries of Monterey Canyon as they appear on Chart III. The mean of the remaining 11 measures is 46.3 feet per mile for the gradient of the shelf as a whole between the shore and contour −600 feet.

Similar measures were made on the charts for the seaward gradient of the Continental Shelf between Pt. Conception and the Mexican boundary. These were of course limited to the slope between the shore and the −300 feet contour, the latter being prevailingly close to the seaward edge of the shelf. A mean of eight measurements showed that in these, for the same spacing as before, the gradients are in feet per mile: 75, 30, 150, 200, 150, 100, 75, 37.6. The mean of these figures is 102.2 feet per mile which is a little more than twice that for the mean gradient for the same part of the shelf to the north of Pt. Sur. This remarkable difference in the seaward gradient of the Continental Shelf north and south of Point Conception is also probably ascribable to the diastrophism which gave rise to the archipelago off the southern coast.

*Dissection of the Shelf.*—The canyons cutting the Continental Shelf are its most important features in regard to the question of the reality of submarine erosion. In the entire length of the 60+ mile belt of sea bottom extending from Lat. 42° N. to Mexico, there are 38 canyons or gorges transverse to the trend of the coast. This number does not include a few small notches in the contours which may be doubtful, and particularly it does not include serrations which are common along certain parts of contours −300 feet and −600 feet. Of these 38 gorges, as I would prefer to call them, there are 23 which are incisive into the Continental Shelf. This does not mean that they cut across the entire width of the shelf. Very few reach the present shore line, and they are much more features of the outer or seaward half of the shelf than of the landward half. A few words descriptive of the dimensions and salient features of some of these may be given here to justify their classification as canyons. All of the 23, which are here segregated from the total number of 38, pursue their course as gorges beyond the outer limit of the shelf to the lower limit of the Steep Slope of Zone II, becoming rapidly larger in cross section in the direction of their down grade. The limit of vigorous downward corrosion is reached when the grade of the bottom of the gorges reaches the top of the embankments which mantle the lower part of the Steep Slope. But some of the canyons, notably Monterey Canyon and Arguello Canyon, are continued seaward beyond the bottom of the Steep Slope in the form of narrow shallow troughs cut below the general slope of the embankments.

These shallow troughs correspond genetically to those which ordinary land rivers not infrequently cut into their deltas after their escape from canyons cut in mountain ridges. On account of this genetic correspondence, the embankments with which we have here to deal are called deltaic embankments.

The small currents, in the small gorges which traverse the outer part of the Continental Shelf—occasionally almost the entire width of the shelf—are clearly engaged in a process of head-water erosion. But the impulse which

generates the current has its origin in every case down stream. That impulse is distributed over a large fraction of the length, and it may be the entire length, of the gorge to the seaward of the shelf. The nature of the impulse is that of the tendency to create a vacuum by slumping of large rock masses, or the suspension and sudden removal of fine material. The direction of removal must always be down the slope of the gorge, that is, of course, down stream. The effect of this process at the level of the shelf is one of suction down into the incipient gorge at the head of the 'canyon'. Once initiated the process would be continuous, making for the progressive enlargement of the 'canyon' in length, width, and depth. The only question that arises is as to the initiation of the process; and the answer to this question is given in the existence of the Steep Slope of Zone II, called the Continental Slope by Shepard and Emery. If the Steep Slope be regarded as a relatively recent upthrust of the sea bottom, as it is by the present writer, the questions both of the initiation and the continuance of the downward corrosive process would be answered. The matter of origin of the Steep Slope will be discussed later in this paper, and we may here return to the features of the Continental Shelf.

*The Shelf from Lat. 42° to Punta Gorda.*—For the region charted by Shepard and Emery, the Shelf extends south from Lat. 42° N. to Punta Gorda with a nearly meridional trend and a mean width of 13.6 miles. At the south end of this stretch of 121 miles, there are three large indentations in the seaward brink of the Shelf which are well shown on the chart, on which they are named Eel Canyon, Mendocion Canyon, and Mattole Canyon. Throughout the stretch, including the three embayments mentioned, there is a pronounced contrast between the slopes above and below the −600-foot contour, which is taken to be the outer limit of the shelf. The slope above that limit is about 28 feet to the mile and is slightly convex upward. Below the same limit and northward of Eel Canyon, the slope across Zone IV is 96 feet to the mile at the latitude of Pyramid Point and also at Trinidad Valley. It is of interest here to note that, for a stretch of about 25 miles between Eel Canyon and Punta Gorda, the top of the Steep Slope rises to and coincides with the −600-foot contour at the brink of the Continental Shelf, so that for this distance Zone IV is entirely eliminated. This elimination of Zone IV also suggests the possibility that the San Andreas rift here becomes the brink of the Continental Shelf, and passes on seaward by way of the lower part of Eel Canyon, thus explaining the complicated configuration of that embayment of the Steep Slope.

Eel Canyon has undercut the Shelf and placed the −600-foot contour close to the −300-foot contour, so that it makes a rather large bite into and below the Shelf. It post-dates the formation of the Shelf. Mendocino Canyon and Mattole Canyon are both tributaries of a gorge which follows the east-west Gorda escarpment. The Mattole tributary moreover appears to be on the immediate extension of the San Andreas Rift which is well exposed with characteristic effects at Shelter Cove. Both these tributaries are shown on Chart IV as indentations cutting across the shelf to the −300-foot contour. It thus appears that the course of the San Andreas Rift, whether it turns to the west and becomes the Gorda fault or turns not at all, but heads on seaward through lower Eel Canyon, must for the present be left an open question.

*The Shelf from Punta Gorda to the Golden Gate.*—Following the coast southerly beyond Punta Gorda, three more gorges are encountered. Two of these cross Lat. 40° N. and come together as one feature at Lat. 40° 15′ N. The more northerly of the two is Spanish Canyon which cuts into the Continental Shelf to its middle part and the other Delgada Canyon which similarly cuts almost entirely across the Shelf. A little farther south is Noyo Canyon which has changed the position of the −600-foot contour, but has not reached the −300-foot contour. These three canyons at their intersection of contour −600 feet appear to follow fault lines having in each case an eastward shift of the northerly fault segment, as noted by Shepard and Emery. At least the seaward edge of the Shelf is jogged a mile eastward on the north side of Spanish Canyon. It is similarly jogged 3 miles where Delgada Canyon makes the same intersection. These jogs in the seaward edge of the Shelf, if they be

faults, have no corresponding features on the landward side of the shelf that can be recognized on the chart.

Following the costal trend southward, the Continental Shelf doubles its width from 5 to 10 miles between Punta Gorda and Pt. Arena, a distance of about 100 miles; in this stretch there are no canyons other than the three just mentioned cutting into it or across it. From Point Arena southward to Bodega Head, a distance of 57 miles, the Shelf again doubles its width to about 20 miles, and there are no canyons traversing it. In the next 5 miles the width of the shelf increases by reason of a jog which, as nearly as can be determined, coincides in course with the normal to the coast drawn on the chart seaward from Bodega Head.

With regard to this jog which so abruptly widens the Continental Shelf at the normal from Bodega Head, and which is interpreted by Shepard and Emery as the effect of a fault, it is interesting to note that the −300 foot-contour is similarly offset about 24 miles farther to the southeast without in any way affecting the position or course of contour −600-feet. The effect of this offset (about 9 miles) is to bring the −300-foot contour much closer to, and parallel with, the −600-foot contour. On a normal to the coast through North Farallon Island the −300-foot contour is only 2 miles from the −600-foot contour and the gradient of the shelf becomes 150 feet to the mile; while on the normal through Bodega Head the distance between these two contours is 13 miles. An uplift of the Point Reyes Peninsula on the San Andreas fault, long anterior to the movement of 1906, might explain the two offsets, one of the −600-foot contour and the other of the −300-foot contour, without later transverse faulting.

Regarding Point Reyes Peninsula as essentially insular, and measuring from the east side of Tomales Bay, the widest part of the Continental Shelf for the whole length of the California coast is at the bulge of the −600-foot contour, between the two offsets described above, and is 35 miles, as measured on the normal through Pt. Reyes. Before leaving this part of the Shelf I call attention to the remarkable serrations which characterize the −330-foot contour at its offset. I have suggested for other localities, where the −300-foot contour is the brink of the shelf, that such serrations are indicative of a crumbling of the brink, or, to put it another way, of a destructive retreat of the front of the shelf shoreward. The serrations continue from the vicinity of Pt. Reyes for 100 miles to Monterey Bay.

From the normal through Bodega Head to that through the Golden Gate, a distance of 46 miles, contour −600-feet is both the seaward edge of the Continental Shelf and the upper limit of the Steep Slope. For this distance Zone IV is entirely eliminated from the submarine profile.

*The Shelf from the Golden Gate to Mexico.*—From Tomales Bay to the Golden Gate the shelf has a width somewhat less than the maximum. From the Golden Gate southward it becomes narrower due partly to a bending of the course of contour −600 feet toward Monterey Bay. This bending is doubtless due to an accommodation of the shore line and of the profile of the sea bottom to the shallow syncline in which Monterey Bay lies. Opposite the shore feature called Pillar Pt., two tributaries of Pioneer Canyon working in unison have made an embayment in contour −600 feet for 3 miles shoreward from the prevailing course of the contour and 6 miles wide in the direction of that course. Ascension Canyon, at Lat. 37° N., makes a sharp indenture of 4 miles into the shelf, and its first tributary to the south makes a similar but smaller embayment.

Following the coast southward on the charts, the next submarine canyon which intersects the shelf is Monterey Canyon and its tributaries the largest and most remarkable of them all. It starts at the shore line of the head of Monterey Bay, where the shelf is 16 miles wide and completely transects that feature as a deeply incisive gorge. Of its two more important tributaries Soquel Canyon on the north side is 13 miles long and entirely within the area of the shelf, while Carmel Canyon on the south side, also 13 miles long, is beyond the brink of the shelf except for the first 2 miles of its course from the shore.

The bottom of Carmel Canyon joins that of Monterey Canyon at a depth of 7200 feet at a point 9 miles from Pt. Pinos. The bottom

of Soquel Canyon joins the same trunk channel at contour −2500 feet. The course of the trunk channel between these two junctions is 14 miles and the gradient between is 335 feet to the mile. To the south of Carmel Canyon, two other tributaries of Monterey Canyon make embayments in the −600-feet contour but do not cut the −300-foot contour. As already stated, the −600-foot contour at Lat. 36° 15′ ceases to be the seaward boundary of the Continental Shelf, and the brink of the latter is close to contour −300 feet from this latitude to the Mexican Boundary. In this stretch, having a length along the coast of about 463 miles, the shelf is cut by 13 'canyons', though nowhere completely to its landward boundary. The mean interval between these is 38.6 miles, whereas north of Pt. Sur to Pt. Arena the mean interval between the canyons is 14 miles. The mean width of the shelf, as attenuated to the −330-foot contour south of Pt. Sur is 4.4 miles; while that north of Pt. Sur to Pt. Arena is about 17.2 miles.

*Summary of Shelf features.*—The Continental Shelf has the characteristic features of continental shelves the world over. In addition, it is noteworthy for two aspects which are perhaps exceptional. One of these is the hypsometric change in its seaward edge from −600 feet to about −300 feet at Pt. Sur in following it on the charts from north to south; a change which may well be genetically related to the crustal movement that gave rise to the archipelago off the coast of southern California. The second exceptional aspect, which may be exceptional only by reason of the writer's ignorance of the submarine features of other coasts, is the extent to which the Continental Shelf of California is traversed by gorges or canyons. These canyons are mostly small, but farther seaward, where they traverse the Steep Slope, they become large and impressive features of the sea bottom.

*The Narrowing of the Shelf and subsidence.*—It has been suggested in a foregoing paragraph that the narrowing of the shelf at Point Sur may be related to the subsidence which gave rise to the archipelago. To clarify this possible relationship it is necessary here to discuss briefly the cause of the sinking of the sea bottom in this region. Such a depression of the sea floor, as is displayed in the physiography of the archipelago, could only occur as a movement consistent with the principle of isostasy. Before, during, and after the subsidence, isostasy, of course, prevailed as a governing principle. In speaking of the compressive stress which characterizes the earth's crust off all continental coasts, I have elsewhere indicated that it arises from a landward inflow from below the ocean to compensate the loss of mass which the continents suffer by erosion. The stress develops to the point of crustal deformation in the more superficial rock riding on the inflow. It accumulates steadily at a variable rate determined by the distribution of continental relief and the vagaries of its reduction. But the relief of the stress by fracture and uplift may be spasmodic, affecting one portion of the coast before another.

Wherever the deep inflow of heavy rock advances far enough to cross the shore line, in one of its spasmodic or local movements, it is added to the vertical section of the continental margin, both as to thickness and mass. The first effect may be an uplift of the land surface; but, since the insertion consists of rock of relatively high density, the area affected will begin to sink, and will continue sinking until isostatic balance is re-established. This vertical collapse may well explain the subsidence which gave us the archipelago with which we are here concerned. That subsidence probably extended southward from Pt. Sur to beyond the Mexican boundary. To illustrate the mechanism of this hypothesis more concretely, we may assume that the mean altitude of the area of subsidence was 100 feet before this particular insertion of the inflow, and that as a result of the depression it was flooded by the sea to a mean depth of 1,000 feet. There was thus added to the column a mass of $1.026 \times 1000 = 1026$ per square foot of area. If $x$ be the mean thickness of the inflow in feet, the added mass at a density of, say 3.05, was $3.05x$ per square foot. To make room for these two additions there was removed from the column $(1100 + x)$ feet at a density of, say 2.67, having a mass of $2937 + 2.67x$ per square foot. To maintain isostatic balance the mass removed was the same as that added, so that $2937 + 2.67x = 1026 + 3.05x$ whence $x = 5029$ feet. This figure is not necessarily the

thickness of the inflow as it approached the coast, but that which it attained by piling up, perhaps unevenly, against the added resistant stress of the landward thickness of the crust as compared with that beneath the sea floor.

The foregoing remarks have for their purpose merely the rationalization of the suggestion that the Steep Slope owes its uplift to the inflow of heavy rock from beneath the ocean to compensate the continental erosional loss of mass. As this inflow enters the coastal belt, it encounters greater resistance owing to the increase of load to be lifted of more than 2 miles of rock, represented by the continental seaward slope.

In the recognition of the origin of the archipelago of Southern California as due to local or spasmodic invasion of the inflow, we have also recognized that it is on its way beneath the continental surface, and that as it proceeds Southern California is due to develop, and perhaps has developed, symptoms of instability, such as Gilluly has reported (1949). We may also infer that in time to come, perhaps not so very far in the future, the inflow is liable to renew its break through into the more northerly continental region. It is not improbable that movements, such as those on the San Andreas rift, are preliminary to such a break through.

### *The Features of the Archipelago*

It may be well here to call attention to the fact that the subsidence which gave rise to the archipelago was not a uniform depression of the sea floor. The latter is characterized, not only by islands but also by basins and submarine ridges, all lying landward of the Steep Slope. These features suggest, if they do not prove, a greatly broken sea bottom, as if the invading inflow were far from being uniformly thick.

The islands of the archipelago are in two groups. The first group comprises Anacapa, Santa Cruz, Santa Rosa, San Miguel, lying on an east-west line at about Lat. 34°, and all four rising from the same broad shelf, the outer limit of which is −600 feet. The second group is situated in the central part of the archipelago, and every island in it rises from its own separate shelf, the outer limit of which is also −600 feet. The group comprises Santa Catalina, Santa Barbara, San Nicolas, and San Clemente. With these we may include Tanner Bank and Cortes Bank which are almost islands, and which also have separate shelves out to −600 feet. We thus have 10 islands, and 7 shelves at −600 feet, and we are concerned here particularly with the shelves and with the hypsometric contrast which they all present to the continuous Continental Shelf at about 300 feet of the mainland shore.

The area of the archipelago apart from the insular features is chiefly occupied by basins together with their upward slopes, and a few submarine ridges. The well defined basins are 11 in number, and their names together with their maximum depths as given on Chart I of Shepard and Emery are: Santa Barbara Basin −2956 feet, Santa Monica Basin −3078 feet, San Pedro Basin −2994 feet, San Diego Trough −2928 feet, Santa Cruz Basin −6450 feet, Catalina Basin −4452 feet, San Clemente Basin −6912 feet, San Nicolas Basin −6012 feet, Cortes Basin −6492 feet, Tanner Basin −5088 feet, Long Basin −6360 feet.

The coincidence of the depth of the outer limit of the shelf of all the islands with the corresponding limit in the region north of Pt. Sur and its contrast with the depth of the edge of the shelf along the shore from Pt. Sur to the Mexican boundary are significant facts. They signify that the islands have not participated in the last depression of the sea bottom, and that the depression was irregular in its distribution, limited to the basins. The absence of a −600 foot shelf off the shore of the mainland from Pt. Sur to Mexico has been explained as due to its reduction to −300 feet as a result of subsidence. But the −600 foot shelf which surrounds all the islands of the archipelago, as a group or individually, indicates that the islands existed very much as they are today at the time of and during the depression. They may of course have acquired their insular character at the time of an earlier subsidence of the sea floor, since it is probable that there has been a long series of such movements.

The basins fall into two groups as to depth. Those of the first group in their elongation follow the trend of the mainland shore, and have on the map the form of canoes strung along parallel to the general trend of that shore. The depth of these four basins ranges from 2928 to

3078 feet with a mean of 2977 feet. The basins of the second group range from 4452 to 6912 feet with a mean of 5966 feet, a little more than twice that of the first group. Their sides are much steeper than those of the first group and are, on one or both of their containing slopes, fairly certainly normal faults; so that they have the characteristics of graben.

The first group of islands, aligned on Lat. 34°, appears to be a seaward extension of Santa Monica Mountains, but none of the other islands of the archipelago are similarly related to the relief of the mainland. The islands of the second group all lie within a radius of about 42 miles from the center of the archipelago, taken to be at Long. 119°, Lat. 33° 10′. But the submarine ridges, whether they culminate in islands or not, are systematically disposed in a general way parallel to the course of the Steep Slope which is the sharp seaward limit to the whole archipelago.

We have thus presented to us, when we scrutinize the physiography of the archipelago, a remarkably irregular and partial, but nevertheless systematic, collapse of the sea floor under the control of isostatic balance. The hypothesis here advanced to account for the phenomena is an inflow of heavy rock in depth from the oceanic area, breaking through the Continental border in irregular slow currents of variable thickness, after the uptilt of the Steep Slope. It would be interesting to check the hypothesis by determinations of the force of gravity at as many points as possible, within the area of the archipelago.

*Zone II—the Steep Slope or Continental Slope*

*General features.*—The most characteristic features of the Steep Slope on the charts are apparent almost at a glance, particularly when the sheets are mounted together as one map. These more striking features are: (1) a relatively narrow, fairly constant width; (2) a remarkably straight course for the latitudinal extent of California and beyond, both ways; (3) offsets by transverse faults; (4) approximate parallelism with the coast north of Pt. Conception; (5) a relatively steep slope, mantled at its base by overlapping embankments; (6) at the south this Steep Slope forms the seaward boundary of the archipelago off Southern California.

The top of the Steep Slope coincides with the seaward edge of the Continental Shelf (contour −600 feet) at three separate parts of its course. These are: (1) for 18 miles northward of Punta Gorda; (2) from Bodega Head southerly for 46 miles to the normal drawn through the Golden Gate; and (3) from the normal drawn through Santa Cruz for 18 miles to Soquel Canyon, a total of 82 miles.

Since the Steep Slope as a measurable feature emerges from beneath embankments, in all sections its full vertical extent remains unknown.

The accompanying tabulation gives the figures for the vertical range, the vertical measure, the width, and the gradient of the Steep Slope at ten localities on its course, as taken from the charts.

The figure given in the tabulation for the mean gradient does not imply that the slope of the Steep Slope is straight. The contours, which express the angle of slope, are not uniformly spaced on most of the vertical sections. On a few sections the slope is a straight line, but on several the contours are notably closer at the bottom of the slope and farther apart at the top. In some the reverse is true; they are closer together at the top and a little farther apart at the bottom. In still other sections, the contours are at their closest in the middle of the slope and wider apart at both top and bottom. All this means that, accepting the contours at their face value, the slope is not straight in any direction for more than a small fraction of its length, but may be straight for a quarter to the whole of its vertical dimension.

The cause of this variation in profile is unknown. It may be a vagary of erosion, or it may be due to buckling of the slab, or to some other unknown cause. The position and orientation of the bottom of the thrust slab are unknown and we have no information as to its thickness, except that the thrust plane is probably roughly parallel to its top, that is, to the Steep Slope.

*The strike of the Steep Slope*[2].—The Gorda Escarpment as mapped on Chart IV appears to be either the continuation of, or a branch from, the San Andreas rift. The latter has been traced

[2] The term Steep Slope as used in this paper is called the Continental Slope by Shepard and Emery. But the real continental slope comprises a wider and more complex belt of the sea floor than this particular feature.

on its northerly course as far as Shelter Cove by the present writer; beyond the cove it appears according to the chart to curve westerly on its strike into the Gorda Escarpment, which appears to be independent of the position of the latter. The bottom of the Steep Slope is 55 miles distant from the shore as measured on Lat. 42° N. The abutment of the same feature

TABLE 1.—DIMENSIONS OF THE STEEP SLOPE AT TEN LOCALITIES

| Locality | Vertical range | Vertical measure | Width in miles | Gradient in feet per mile |
|---|---|---|---|---|
| Lat. 42° | −3,600′ to −9,900′ | 6,300′ | 10 | 630′ |
| Lat. 40° 30′ | −600′ to −7,500′ | 6,900′ | 10 | 690′ |
| Lat. 39° | −3,600′ to −10,500′ | 6,900′ | 10.5 | 657′ |
| Lat. 38° 30′ | −1,500′ to −11,100′ | 9,600′ | 21 | 457′ |
| Lat. 38° | −600′ to −10,500′ | 9,900′ | 14 | 707′ |
| Lat. 37° 07′ | −3,000′ to −9,000′ | 6,000′ | 10 | 600′ |
| Lat. 36° 15′ | −6,000′ to −10,500′ | 4,500′ | 7.5 | 600′ |
| Lat. 35° | −6,600′ to −12,900′ | 6,300′ | 11 | 572′ |
| Lat. 33° | −5,100′ to −12,300′ | 7,200′ | 8.5 | 847′ |
| Lat. 31° 50′ | −5,700′ to −12,000′ | 6,300′ | 7 | 900′ |
| | | Mean height 6990 ft. exposed to sea water | Mean width 10.95 ml. | Mean gradient 666′ per mile |

trends nearly east-west (N. 87° W.) and so dislocates and offsets the features of the off-shore sea bottom here comprised in Zones I, II, III and IV. The region to the north of Gorda Escarpment seems to have suffered no horizontal displacement. The Steep Slope from Lat. 42° N. to Mendocino Canyon is ideally regular and undeformed, and all the horizontal displacement of the fault movement appears to have been to the south of the great east-west escarpment. The *relative* horizontal shift is the same in direction as that which is manifest along the San Andreas rift for its entire length. There is, however, to the north of the escarpment a vertical downthrow of about 4,500 feet, the equivalent of which is as yet unknown on the San Andreas rift. It is probably the vertical element of the movement on the Gorda Escarpment which gives it such great prominence as a feature of the sea floor off Cape Mendocino. The measure of the horizontal shift is probably not in excess of 50 miles, while the length of the scarp is probably over 100 miles.

The remarkably straight alignment of the Steep Slope for 115 miles, from Lat. 42° N. to Mendocino Canyon has a course of S. 13° E. The trend of the corresponding stretch of the present shore line is S. 5°W. The angle between the course of the Steep Slope and of the shore is thus 18°; and the orientation of the former upon the Gorda Escarpment is 21 miles due west of the shore.

The sea floor south of the Gorda Escarpment down to the vicinity of Pt. Arena presents an astonishing contrast to that on the north side in the orientation of its relief. We have, however, the same four zones: I. The Continental Shelf with a width of 13 miles just north of Pt. Arena, but tapering to nothing 11 miles west of Punta Gorda; II. The Steep Slope, rising from −10,500 to −6,000 feet notably steeper from Pt. Arena to Fort Bragg than it is farther to the northwest, where it is offset seaward on the south side of the Gorda Escarpment; III. The deep sea floor mantled by a well defined large deltaic embankment; and the gentle but unsystematic slope of the non-persistent Zone IV. The latter is wider than is usual north of Pt. Conception and is traversed by three notable canyons named Spanish, Delgada, and Noyo, all of which reach back into the Continental Shelf.

From the end of Gorda Escarpment as mapped at the western limit of Chart IV, the course of the Steep Slope is geographically a chord of the curve which is well displayed in the course of the escarpment and thence in the shoreline from Punta Gorda southward to Pt. Arena. The chord, however, is also slightly concave seaward. There is here at least a geome-

trical relation between the course of the Steep Slope and that of the shore, a relation which doubtless inheres in the fault displacement that gave rise to the remarkable Gorda Escarpment.

From the bottom of Chart IV, Lat. 39°, nearly to Lat. 37° 30′, the middle line of the Steep Slope is straight with a course of S. 25° E. But the landward side is slightly convex as far as Bodega Head and the seaward side is irregular. The irregularity, however, is due to the high parts of the embankments running up into the gorges which dissect the slope; and the convexity of the landward side may be due to an underlying surface of marine erosion.

Beyond the normal to the coast at the Golden Gate, the contour −600 feet continues southerly, with no important change of course or of definition, as the seaward boundary of the Continental Shelf. But that contour ceases somewhat abruptly to be the upper limit of the Steep Slope. Indeed, beyond that normal, and below the shelf, there appears to be a change in the gradient of the Steep Slope into that of Zone IV, and there is no sharp line of separation for a few miles between the two types of slope. Below the Continental Shelf, the whole slope appears to be offset by a jog of 9 or 10 miles, as measured on the abrupt shift of contour −10,500 feet, whereby the zone of the Steep Slope, but not the shelf, is thrown westward. This fault if followed landward would cut the shore just south of Pillar Point. It thus lies to the south of the granite near Pillar Point and also to the south of the granite of the Farallon Islands. About 15 miles south of the offset is the main mass of the Pioneer Seamount lying in the zone of the Steep Slope, but with a typical part of the latter lying to the west of the seamount and having its regular trend. About 25 miles south of the Pioneer Seamount is another, the Guide Seamount, which lies west of the course of the Steep Slope. From a point midway between the two seamounts, the Steep Slope changes its course from S. 26° E. to S. 56° E. and continues on the changed course for 82 miles to Monterey Canyon. Beyond Monterey Trough to the south limit of Chart III, the zone of the Steep Slope resumes its normal course of S. 26° E. for something over 30 miles. It is thus apparent that between this last mentioned portion of the Steep Slope and that which turned in toward Monterey Bay, with the abnormal course of S. 56° E., there is either a large dislocating fault or there are two steep slopes in Zone II meeting in Monterey Canyon, one from the north and one from the south, a possibility which is discussed in the next Section dealing with the origin of the Steep Slope. The 30-mile stretch of the zone of Steep Slope extends to the mouth of Sur Canyon at the southern limit of Chart III, and its vertical range is between contours −6,000 and −10,500 feet.

From the mouth of Sur Canyon at Lat. 36° the upper part of the Steep Slope continues southward with a course of S. 25° E. for 78 miles to Lat. 35°. The lower part of the slope is here concealed by an overlapping complex embankment which raises the sea floor at the outlets of Sur and Lucia Canyons, and which surrounds the Davidson Seamount. At Lat. 35° the zone is cut by a fault which offsets it about 10 miles and marks a change in its course to S. 48° E. This changed course persists, along the Santa Lucia escarpment for 72 miles to Lat. 34° 10′. Near its southern end, however, the escarpment is interrupted by the wide Arguello Canyon and its course thereby somewhat changed. If the dip of the fault at Lat. 35° be southerly it may intersect the shore at the south side of Estero Bay, where there is a similar topographic offset. Between Lat. 34° 10′ and Lat. 33° 50′ the continuity of the Steep Slope is interrupted for a space of 24 miles by the Rodriguez Seamount, which is almost surrounded by an embankment at least 3,000 feet thick.

South of Lat. 33° 50′ the Steep Slope is well displayed on Chart II, under the name of Patton escarpment, for a distance of 63 miles to the bottom of the Chart at Lat. 33°. The trend for this stretch has the normal course of S. 25° E. From Lat. 33° to the southern limit of Chart I the Steep Slope is a very pronounced feature called the Patton Escarpment, the course of which is S. 34° E. It is traversed by three wide canyons at Lat. 35°, 32° 10′ and 31° 50′, all of which head into the most westerly submarine ridge of the archipelago. There is a longitudinal fault at the limit of the chart.

From Lat. 33° to Lat. 31° 30′, the Steep Slope is well developed both as to its expression in contours and general straightness on a course

of S. 29° E; but the top of the slope is lowered from −3,600 to −6,000 feet, with little or no change in the bottom of the Steep Slope at −12,000 feet, and it has a local seaward bulge nearly 30 miles long to about Lat. 32° 20′. The bulge is asymetric and is situated just north of a large canyon which traverses the belt and jogs the contours seaward about 7 miles. The bulge in the otherwise straight course of the belt may thus be due to a fault offset.

*Origin of Steep Slope.*—Having here completed my descriptive remarks regarding the features, course, and complications of the Steep Slope, I proceed to discuss briefly the origin of this remarkable phenomenon of the sea bottom. I have elsewhere in this paper given reasons for the rejection of the notion of Shepard and Emery that the Steep Slope, or Continental Slope as they call it, is a fault with seaward dip, and I here offer another hypothesis of its genesis not so inconsistent with the doctrine of isostasy.

The classical study of the denudation of the area of the United States by Dole and Stabler (1909) indicates that the load removed to the sea by river erosion, at the present time, is that of a layer of rock over the entire area having a thickness of one inch, in 760 years. In the time since the close of the Cretaceous, about 55,000,000 years, at this rate, over a mile of rock would have been removed from the area of the United States and delivered to the ocean. That the rate of erosion was the same as at present throughout the Tertiary is of course questionable. But we may infer that the erosional removal since the end of the Cretaceous, judging from the record of marine deposition, was at least a large fraction of a mile thick over the area of the United States.

If this be so then, as between the continental and the oceanic areas of the globe, there is, and always has been, a tendency to develop a state of imbalance. But the counter tendency to restore balance is no less constant. In so far as is known to geology there are two, and only two, ways by which isostatic balance may be maintained as between land and sea. One is a deep inflow of heavy rock from beneath the sea to the land to compensate for the loss of mass by erosion; the other is a general deformation of the geoid, whereby the sea bottom sinks and the land rises. It is possible that both of these mechanisms may be continually simultaneous, but it is highly probable that one or the other is the chief agency making for the restoration of balance. Since Barrell's discovery of the asthenosphere, we may confidently assume that a landward flow, in that relatively mobile layer, across the continental margin, is the chief mechanism. A general effect of that flow is that the more rigid overlying crust tends to ride with it. Hayford (1911) clearly recognized the riding of the upper crust upon the compensatory inflow. His words are:

> "Horizontal compressive stresses in the material near the surface above the undertow (inflow) are necessarily caused by the undertow. For this undertow necessarily tends to carry the surface along with it and so pushes this surface material against that in the region of erosion."

The general effect of this pushing of the crust riding on the inflow is that of thrusting with its concomitant diastrophism. One effect of the compressive stress is, as I have already pointed out, the inhibition of normal faulting, except in directions approximately normal to the compressive stress; and the effect of thrust is uplift, such as that displayed in the Sierra Nevada.

In an inquiry as to the genesis of the Steep Slope, the effects just mentioned must be taken into consideration. A normal fault being ruled out and uplift very apparent it appears that we are limited to thrusting as the principal causal factor in the genesis of this astonishing feature of the sea bottom. The slope itself is apparently the uptilted surface of the sea bottom, but this implies either a break in the sea bottom to the east of the Steep Slope, or a reversal of slope to the east, as in the case of an an anticlinal bulge. But in the contours of the charts there is little or no evidence as to whether the Steep Slope is the seaward flank of an uplifted thrust block, or the west flank of an anticline, the east flank of which is buried beneath the formations of zones IV and I. This is particularly true to the north of Lat. 35° 30′, but to the south of this latitude there is a submarine ridge which extends through the archipelago nearly to the southern limit of Chart I, a distance of 300 miles. This ridge includes Santa Lucia Bank, the islands of San Miguel and Santa Rosa, and the Santa Rosa-Cortes Ridge,

Bishop Rock, and Sixty Mile Bank. It lies from 20 to 50 miles shoreward from the top of the Steep Slope and may possibly be significant of an anticlinal axis extending through zone IV. Analogy with the structure of the Sierra Nevada is far more suggestive of an uplift on a thrust fault.

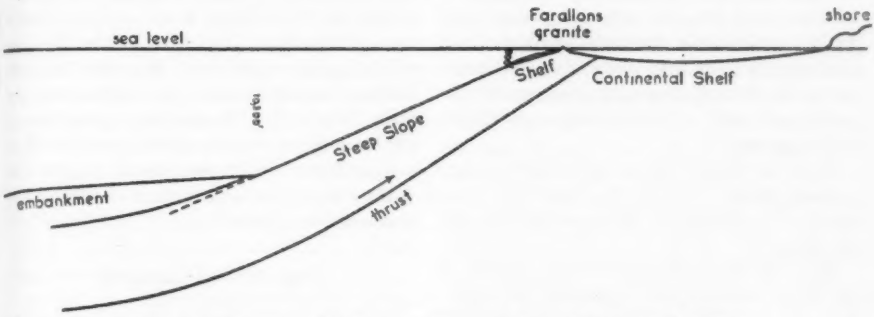


FIGURE 1.—HYPOTHETICAL SECTION THROUGH AREA OF FARALLON ISLANDS

In this paper I have recognized the Steep Slope as an uptilted zone of the sea bottom having, as a result of the tilt, a mean seaward down gradient of 666 feet per mile. I have also called attention to the fact that the western limit of this slope is everywhere buried beneath depositional embankments. The regularity of the slope as to its course and the approximate constancy of its gradient suggest constancy of petrographic character and that its surface before the uptilt was that of an even sea floor, devoid of relief. This evenness of surface might well be supplied by a sedimentary formation which had been deposited on the sea floor without appreciable dip. But such a body of sediment, unless it were limestone, would, if tilted, have little resistance to mechanical disintegration, with corresponding development of relief. In the stretches between the numerous gorges or 'canyons' which transect the Steep Slope, however, its surface, insofar as can be judged from its representation in contours, is astonishingly even and free from the detail of relief. It appears to be composed of a hard, resistant rock, and the walls of the gorges which transect it have also relatively steep slopes. It may well be granite. There are many extensive regions of the earth's surface composed of granite, which are geomorphologically peneplains, and locally plains. So far as I am aware such features may not be excluded from the sea floor off continental margins, in our present state of knowledge.

The notion that the Steep Slope may be underlain by granite is supported by at least one fact of observational value. That is the position of the granite ridge of which the Farallon Islands are the culminating peaks. The ridge is at least 8 miles long and its crest is parallel to the west face of the Steep Slope at contour −600 feet. Here, opposite the Golden Gate, near the outer edge of the Continental Shelf, and 4 miles seaward of the Farallons, is the submarine Farallon Escarpment of Chart III. The top of the Escarpment is a culminating line of the Steep Slope. If the surface of the latter be an uptilted granite surface the ridge of the Farallon Islands may well be the culmination of the uplift. Immediately to the east of the islands would lie, under the shelf embankment, the outcrop of the thrust fault upon which the Steep Slope block considered as a diastrophic feature, was shoved eastward and so uptilted. These hypothetical relations are indicated in the accompanying Figure I in which the vertical scale is 5 times the horizontal. The angle of dip of the thrust fault is of course unknown. The hypothetical fault noted as extending from Punta Gorda to the lower Eel Canyon may possibly be the same thrust which is here suggested as probable to the east of the Farallon Islands, beneath the embankment of the Continental Shelf. The evidence for this supposition is, however, not entirely satisfactory.

The coincidence of the top of the Steep Slope

with the seaward edge of the shelf at contour −600 feet, from about latitude 37° to Soquel Canyon, appears to be the southward extension of the relations just mentioned for the latitude of the Farallon Islands. The granite north of the city of Santa Cruz, although it is surrounded by Miocene sediments, would then be the culmination of the upthrust which tilted the Steep Slope. A field study of the northeast side of the Santa Cruz granite might contribute to the verification, or the negation, of the thrust here suggested.

On Chart III of Shepard and Emery a fault is shown on the northeast side of the Santa Cruz granite, but whether or not it is a thrust is not apparent.

Monterey Bay lies at about the middle of the entire extent of the Coast of California. Thus far, in an effort to interpret the origin of the Steep Slope, I have assumed that the latter is a single feature from Oregon to Mexico. It is here a convenient place in the discussion, to question the validity of this assumption, or rather to present another point of view as to the unity of the Steep Slope, considered as a geomorphic feature. The disposition of the Steep Slope in the vicinity of Monterey Bay as indicated on Chart III, would perhaps justify the supposition that this great feature of the sea bottom represents two geographically distinct movements of the compressive stress to which both owe their existence. In other words it may be a better statement of the case to say that there are two stucturally identical features having the characteristics of the Steep Slope, as the term has been here used. These may be referred to as the northern Steep Slope and the southern Steep Slope. The two meet in Monterey Bay and the gorge of the submarine Monterey Canyon is the place of meeting.

From this point of view the pronounced change of course of the northern Steep Slope, at the interval between the Pioneer Seamount and the Guide Seamount, whereby it heads into Monterey Bay, signifies approach to the end of that feature of the sea bottom. If the changed course continues beneath the embankment of the Continental Shelf in Monterey Bay, it will intersect the thrust fault which I have hypothetically placed on the north-east side of the Santa Cruz granite. Such an intersection would bring to an end the northern Steep Slope.

The southern Steep Slope, on its northerly course, has very closely the same bearing as that displayed by its counterpart north of Pioneer Seamount. But to the north of its intersection by Sur Canyon, it too curves easterly toward Monterey Bay. The gentle syncline which comprises Monterey Bay thus becomes common to both the north end of the southern Steep Slope and to the south end of the Northern Steep Slope. And the granite, which ends as a shore feature at Pt. Sur, would become the uptilt edge of the same surface as that of the southern Steep Slope.

### *Zone III—Embankments*

Zone III of the relief of the sea bottom has for its dominant characteristic the overlapping embankments which extend seaward from the lower part of the Steep Slope. As to origin, these are of two kinds. The first comprises those embankments which surround the seamounts, of which five are distinguished by name on the charts. The relation of these encircling embankments to the seamounts agrees with the suggestion that the latter are extinct or dormant explosive volcanoes, and that the deposits are volcanic ejecta. The second kind are related in position and in slope of surface to the canyons which traverse the Steep Slope. The highest point of their individual surfaces is, in every case, at or near the mouth of a canyon, and in this respect the embankments resemble the alluvial fans which have been built up at points where rivers emerge upon low lands by way of gorges from hilly or mountainous country. The lower part of the Steep Slope emerges from beneath these embankments, so that nowhere can its actual bottom be determined. The lower part of most of the embankments extends seaward beyond the western limit of the charts. In several cases, the embankment extends up into the canyons. But lower down the slope it may be difficult or impossible to discriminate an embankment built up around a seamount from one originating at the mouth of a canyon.

A large embankment of Zone III lies immediately north of the Gorda Escarpment. It ap-

pears to be due to the combined discharge from three canyons named the Eel, the Mendocino and the Mattole. The apex of the slope heads well up into Mattole Canyon. The top of this embankment has an altitude of −7,500 feet and the deepest water recorded in contours north of the escarpment is −9,900 feet. The configuration of its surface is clearly indicated by contour −9,000 feet. The gradient in the middle part of the slope for the first 46 miles is 130 feet to the mile, and for the first 65 miles it is 100 feet to the mile. The total slope to the −9,900 foot contour is thus concave upward. The present depth of the original sea floor upon which the embankment was first deposited is unknown. However, several other embankments slope down below −12,000 feet, and it is probable that the original sea floor is now deeper than this. Assuming this to be so we may arrive at a probable figure for the minimum thickness of the embankment lying north of Gorda Escarpment. Where it is traversed by contour −9,000 feet, it has a probable least thickness of 12,000 feet −9,000 feet = 3,000 feet. Up the slope the contours above −9,000 feet indicate some addition to this estimate.

Trinidad Valley, about 35 miles north of Eel Canyon has a gradient of 244 feet to the mile, and a well defined deltaic embankment. The gradient down the slope of the latter is 43 feet to the mile. The lower limit of the embankment is beyond the boundary of the chart.

In the latitude of Fort Bragg there is a pronounced embankment outlined by contour −10,500 feet due to deposition from Delgada and Noyo canyons. Its breadth normal to the coast is 36 miles within the contour mentioned. Its thickness at contour −10,500 feet is probably 2,400 feet. Its gradient on Lat. 37° 42′ out to contour −12,000 feet is 80 feet to a mile.

The next canyon to the south is unnamed. It cuts through the Steep Slope at about Lat. 37° 30′. The embankment in front of this canyon is outlined by contour −12,000 feet, and its breadth within this contour is about 27 miles. The gradient of this embankment is 53 feet to the mile. The Pioneer Seamount is situated just south of the mouth of the canyon and there may be some volcanic ejecta in its make-up. Pioneer Canyon cuts through the Steep Slope between Pioneer and Guide Seamounts, and although there is an embankment at its mouth, outlined by contour −10,500 feet, there is probably a large amount of volcanic material in it.

Monterey Canyon has a relatively large embankment which extends northward to a mergence with the embankment of Pioneer Canyon and south to a mergence with the embankment which surrounds the Davidson Seamount. It also slopes well up into the gorge of Monterey Canyon. Seaward this embankment after having been built from a level below −12,000 feet to above −10,500 feet, was again dissected to −12,000 feet. It seems probable that the upbuilding of the sea floor was due to an excess of ashes delivered to the region by the volcanic activity of the Davidson Seamount, and that when this supply diminished later erosion made a trench across the embankment.

Sur Canyon to the north of Lat. 36° and Lucia Canyon to the south have a common deltaic embankment. It extends seaward for about 12 miles from the bottom of the Steep Slope with a down gradient of 56 feet to the mile toward the larger embankment which surrounds Davidson Seamount. It has also a westerly downward slope toward the embankment of Monterey Canyon, and merges with it.

Arguello Canyon has a notable embankment, the westerly gradient of which above Lat. 34° 10′ is about 53 feet to the mile. The much larger embankment south of latitude 34° lies seaward of Rodriguez Seamount and is probably chiefly composed of volcanic ejecta. But there must be in its make-up a large admixture of detritus from Arguello Canyon. From Rodriguez Seamount southward the sea floor, west of the Steep Slope, appears to be infested with small seamounts, so that the character of the embankment cannot be determined. One more canyon with an embankment extending seaward from its mouth may be mentioned. The canyon has an east-west course in Lat. 31° 45′, but the embankment extends seaward to the southwest. Its gradient for 17 miles seaward from the foot of the Steep Slope is 35 feet to the mile.

*Zone IV*

The areas which are here classified as zone IV of the charts remain for descriptive comment. The zone is not a continuous belt of the sea floor. There are sections of the latter in which there is no representation of the particular spacing of contours which characterizes the zone. The only other characteristic besides the angle of declivity is that of geographical position. The areas comprehended in Zone IV lie, as the charts well show, between the flat, lowly inclined Continental Shelf and the Steep Slope.

The most northerly of these areas extends as a belt of the sea floor from Lat. 42°, with a fairly uniform width of 24 miles, as far as Trinidad Head, a distance of 66 miles. From this latitude the belt tapers regularly for 38 miles to nothing at Lat. 40° 30′. The tapering appears to be due to a southerly increase in altitude of the top of the Steep Slope to −600 feet, and a westward migration of the outer edge of the Continental Shelf. For 16 miles south of Lat. 40° 30′, the outer edge of the Continental Shelf and the shoreward edge of the Steep Slope have a common boundary at −600 feet, eliminating Zone IV. The southward tapering of Zone IV, and the local elimination of that zone for 16 miles, are probably due to a fault movement.

South of the Gorda escarpment, at the end of the 16 mile interval, we enter again upon an exceptionally wide area of Zone IV which extends in a swelling curve from the western limit of the escarpment as mapped to Pt. Arena, and there closes again to a narrow belt. This widens gradually till at the normal from Bodega Head it is 7 miles across, where the top of the Steep Slope at contour −600 feet again becomes the seaward edge of the Continental Shelf. At this point Zone IV again ends abruptly. For the next 46 miles there is no representation of Zone IV on the chart. For this stretch the Continental Shelf and the Steep Slope have again a common boundary. The end of the 46-mile stretch is at a nearly east-west fault trace which extends seaward from Pillar Point to the north flank of Pioneer Seamount. This fault trace becomes the the northern boundary of another area of Zone IV. This area is triangular in shape and the southern side of the triangle rises from −6,000 feet on the flank of the seamount to the outer edge of the Continental Shelf at 4.5 miles north of the normal from Santa Cruz. The length of the triangle measured as a straight line on contour −600 feet is 60 miles. From the Santa Cruz normal to Carmel Bay, the sea bottom has been deeply dissected by Monterey Canyon and its tributaries. For this stretch of sea bottom spanning Monterey Bay for 33 miles, the erosional removal has been so deep and so extensive that no remnants of Zone IV can be recognized on the chart. South of Monterey Canyon, however, Zone IV is broadly developed in its normal position between the Continental Shelf and the zone of the Steep Slope. Continuing into the next chart, the zone, broadening southward, extends beyond Pt. Conception and merges with the sea floor of the archipelago. The latter like Zone IV lies between the Continental Shelf and the Steep Slope. At the place of mergence of Zone IV with the archipelago the zone is 55 miles wide including in that measure Rodriguez Seamount. At latitude 35° it is broken by an east-west fault which also offsets the Steep Slope to the west of it. From this fault the zone (IV) continues south to Pt. Conception with a width of about 50 miles.

Beyond Pt. Conception Zone IV widens greatly and cannot be distinguished from the sea floor of the southern archipelago, which, with a maximum width of 150 miles, extends south into Mexican water. This mergence of Zone IV with the sea floor of the archipelago, within the same east and west boundaries, leaves little doubt but that Zone IV and the archipelago have a common genesis and geological history. That is, we must entertain favorably the hypothesis that Zone IV and the archipelago are one and the same feature between the Continental Shelf and the Steep Slope. The hypothesis accentuates the time relations of the origin of the various zones of the off shore sea bottom, as pictured in the four charts. Leaving the seamounts, as well as the submarine canyons, and embankments, out of account, it makes the Steep Slope, the last structural feature. That is, it is a diastrophic modification of the sea bottom.

In conclusion, the writer extends his congratulations to the authors of Special Paper 31

on the publication of the charts. These comprise the best picture of the large features of the sea bottom that we have.

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# CENOZOIC MARINE CLIMATES OF THE PACIFIC COAST

By J. Wyatt Durham

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## Abstract

The data available from the regions surrounding the Pacific Ocean during the early Tertiary indicate that the poles and continents could not have been in the positions postulated by Wegener, DuToit, or Grabau, but were in approximately the same positions as at present. During the Paleocene, the 20°C. marine isotherm was north of 49° N. latitude, by middle Oligocene time it had started to shift south of that position, by middle Pliocene it had approached its present position. During the Pleistocene it oscillated both northwards and southwards.

## Introduction

Nearly 30 years have elapsed since Smith (1919) wrote his classic paper "Climate Relations of the Tertiary and Quaternary Faunas of the California Region" and much new information bearing on past climates of the Pacific Coast has accumulated. Some of the data, for instance that on the Plio-Pleistocene of the San Pedro, California, area, has been interpreted in markedly divergent manners (Smith, 1919; Crickmay, 1929; Woodring, Bramlette and Kew, 1946). Nevertheless it is believed that the data now at hand permit of more precise conclusions at many points than those of Smith, and that the major climatic episodes of the Cenozoic of the Pacific Coast can be outlined with certainty, even though there are local disagreements.

In recent years, the work of Chaney (1940; 1947) and his students (especially Axelrod, 1948) in Paleobotany has contributed much to a clearer understanding of the Cenozoic terrestrial climates of the region and their conclusions are in general accord with the data bearing on the marine climates.

Important papers other than Smith's, in which significant data on the Pacific Coast climates appear, are the following: Clark and Vokes (1936), Durham (1942a; 1947), Grant and Gale (1931), Vaughan (1940), Vokes (1940), and Woodring, Bramlette, and Kew (1946). Other papers in which important data are recorded are: Bailey (1935), Bentson (1943),

Crickmay (1929), Durham (1941; 1942b; 1943; 1944), Miller (1940), Nomland (1916a; 1916b; 1917a; 1917b), Vaughan (1917), Weaver (1943), and Wells (1940). In the present paper, the evidence (in part unpublished as yet) afforded by the past distribution of corals, particularly the hermatypic (Vaughan and Wells, 1943, p. 55–56) or "reef corals", has been given greatest weight, while the distribution of other organisms has been considered secondarily. However, in the absence of hermatypic corals, the evidence afforded by assemblages of other organisms has been considered of nearly equal importance.

## Principles

In paleontology it has often been stated that the "Present is the clue to the Past". This is obviously a sound working hypothesis, yet it must be tempered with caution and common sense in its application. For instance, the pelecypod genus *Patinopecten* has been considered a northern ["cold water"] indicator, yet in the Plio-Pleistocene of the Gulf of California region (Carmen Island, etc.), it occurs in strata which also carry hermatypic corals (*Pocillopora*, *Porites*, *Psammocora*). In this case, it is obvious that *Patinopecten* does not represent *cool* water, but rather a southward extension of range of a genus whose members are largely northern in distribution but not necessarily indicators of "cool water". As a general rule, it may be said that within the limits of temperatures normally prevailing in marine waters *stenothermal organisms are in general much more critically limited by minimum temperatures than by maximum temperatures*. As examples of this rule, there are the numerous tropical shallow-water genera of gastropods (*Conus*, *Cypraea*, *Ficus*, *Oliva*, *Strombus*, etc.) which drop out of the faunas as one leaves the tropics, while on the other hand there are equally numerous examples of boreal shallow water genera of molluscs which are found in deep water (lower temperatures) nearer the tropics (*Yoldia*, *Nucula*, *Lima*, *Fusitriton*, etc.). (*See* Hesse, Allee, and Schmidt, 1937, p. 264; and Sverdrup, Johnson and Fleming, 1942, p. 843–873, for discussions on the causes and effects of temperature on distribution.)

A second rule is that in general *the past ecological requirements of associations of organisms are similar to those of similar Recent associations*. There is abundant proof of this premise during the Cenozoic. Tertiary faunas from the East Indies, or Ecuador or Peru, areas which have always been in the tropics (excluding "Continental Drift") show marked similarity (excluding extinct races and species differentiation) to the tropical faunas now living in adjacent seas and clearly indicate that there has been no change in the ecological requirements of those faunal assemblages during the Tertiary. Similarly no one can doubt that the early Tertiary Tethyan faunas of southern France and Italy, with their abundant tropical molluscan genera, reef corals, and large foraminifera, lived in a tropical environment, even though their geographic position is well north of the present day tropics. On the other side of the world, the lower Eocene faunas of the Crescent formation in the state of Washington (at about 48° N. lat.) include reef corals, large foraminifera, brachiopods, and tropical molluscan genera, clearly indicating that this fauna too lived in a tropical environment. Although not stated as such, this premise is clearly implied in Vaughan's (1940) paper on the "Ecology of Modern Marine Organisms with Reference to Paleogeography". Scepticism has often been expressed regarding the validity of projecting the ecologic requirements of living organisms back through geologic time and assigning them to ancestral or related forms. However, when the earlier form is always found in similar (excluding extinct members of the fauna with no living relatives) associations to those in which its living relatives occur, there appears to be no cogent reason to doubt the validity of such assignments.

## Continental Drift

In any attempt to evaluate past climates of the earth, one must consider the validity of the various "Floating Continent" hypotheses, for one of the postulates common to the three most widely accepted versions of this theory is that the continents wandered about with respect to the earth's poles during geologic time. In consequence, the land masses would have had different relative positions to the poles at succes-

sive intervals, and the climatic belts would similarly have occupied different parts of the continental masses at corresponding intervals.

For the purposes of this study, the three most widely utilized modern versions of "Continental Drift" or "Floating Continents", those of Wegener (1923), Grabau (1940), and DuToit (1937), may be evaluated by means of the same lines of evidence. Insofar as the paleogeography of the Tertiary and the distribution of climatic zones during this time are concerned, the general effect of the theories is the same. Even though Grabau has the position of his polar ice caps reversed during the Eocene and Miocene from those of Wegener, the positions of the continents are about the same and the effect of the polar climates on the northern hemisphere faunal migrations and relationships would be the same. Although DuToit has his independent northern and southern land masses, Laurasia and Gondwana, he (1937, p. 283) notes that during the Tertiary the polar shifts worked out by Koppen and Wegener are thoroughly acceptable to his version. Though the evidence presented here is directed specifically against Wegener's "Continental Drift", it is equally applicable to Koppen and Wegener's later modifications, Grabau's "Polar Hypothesis", and DuToit's version of "Continental Drift".

All the above hypotheses, to be workable, need to have the North Pole during different geological epochs in a different relative position to the land masses than at present, and have indicated that during the later Mesozoic and Tertiary this position would be somewhere in the North Pacific. For instance, during the Eocene, Wegener (1923, p. 109) would place the North Pole at 45° N. latitude, and 180° W. longitude, and at the same time have northeastern North America in juxtaposition with western Europe. With this setup, the North Pole is in the northern part of the Pacific Ocean, and the Arctic areas would spread out from there (Fig. 1). Kamchatka and northeastern Siberia would fall within the Arctic Circle, while Alaska and the Aleutian Islands would be within the middle temperate zone or in a position corresponding climatically to present-day Northern California and Southern Oregon.

During the Oligocene, Wegener would have the North Pole at 58° N. latitude and 180° W. longitude, and in the Miocene at 67° N. latitude and 172° W. longitude, so that at any time from the beginning of the Eocene on through the Miocene, Kamachatka would be well within the Arctic Circle and at times within 15° of the pole. Under such conditions, the Kamchatkan fauna should at all times be boreal in character, and during the lower Tertiary at least should have a markedly different character from the fauna of California, which would have been within the tropics. Yet there is a marked similarity between the faunas reported (Ilyina, 1939; Pleshakov, 1939; Slodkewitch, 1938a; 1938b) from Kamchatka and those of the western United States.

During the Eocene-Oligocene of Kamchatka (the faunas of which are apparently poorly known), such species as *Turritella uvasana*, *Turritella* cf. *uvasana*, *Turritella* cf. *blakeleyensis*, *Macrocallista pittsburgensis*, and *Crassatellites* aff. *washingtoniana* of the Pacific Coast faunas are reported as occurring along with other less significant identical or related forms of our faunas. The genera *Turritella*, *Macrocallista*, and *Crassatellites* are typical of tropical to warm temperate faunas and their presence in the Kamchatkan faunas indicates that those faunas lived in warmer waters than those of the present day adjacent seas, not colder as Wegener would have the area. When this fact is considered along with the other identical or closely related species that are found in the Kamchatkan Territory and the Pacific Coast Tertiary, it appears highly improbable that the two regions were as markedly different climatically, or as distant geographically as if the continents were in the positions Wegener postulates.

Further evidence against Wegener's hypothesis is found in the character of the Eocene faunas of the Pacific Coasts of North and South America. As shown later, there is good evidence that the Eocene faunas of the Pacific Coast as far north as 49° N. latitude are of tropical aspect. Along the western coast of South America, the well-known Eocene faunas of Peru (Olsson, 1928; 1929; 1930; 1940) are also obviously tropical in character with various strombid gastropods, species of *Conus*, *Cypraea*, *Ficus*, and *Spondylus*, orbitoidal foraminifera, and various corals. Thus, during the Eocene, there were

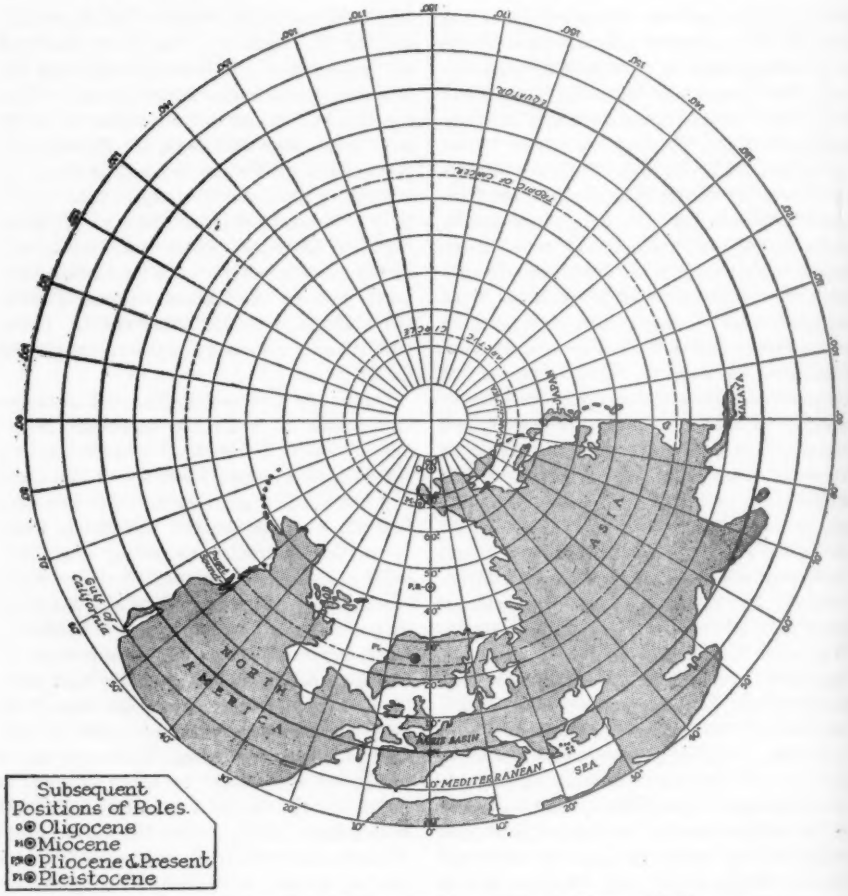


FIGURE 1.—EOCENE NORTH POLAR PROJECTION ACCORDING TO WEGENER'S HYPOTHESIS

tropical and subtropical faunas stretched over 60° of latitude along the Pacific Coast of the Americas, with the southernmost (Peru) some 20° south of the tropics (assuming climatic belts of about the same width as today), if Wegener's position (or anything approaching it) of the poles and continents is accepted. At the same time in Java, on the other side of the Pacific, the Eocene fauna of Nanggulan (Martin, 1914; 1931) is also very obviously tropical in character and must have lived in waters very similar to those now prevailing in its vicinity, less than 10° from the equator. In summary, the tropical and subtropical character of the Eocene faunas surrounding the Pacific basin indicate that the tropical belt of Eocene time must have included all the present-day tropics, and at the same time must have been considerably wider in order to account for the distribution of tropical and subtropical faunas on the Pacific Coast of North and South America and in the Western Pacific. It appears impossible to account for their distribution if the continents and poles were in the positions postulated by Wegener for the Eocene, and with climatic belts of the same character and width as those now prevailing.

An additional bit of evidence against Weg-

ener's postulates is supplied by the relationships of the marine Eocene faunas of the East Coast of North America and of Western Europe (England, France). According to his map (Wegener, 1923, p. 6, Eocene), there would be only about 1,000 or 1,200 miles between the New Jersey-Maryland Eocene and that of Western Europe, with a continuous shoreline, facilitating easy migration, between the two areas. Under such conditions, and with only about 5° difference in latitude (according to his map) between the two areas, the faunas should have at least a moderate number of species in common. Yet, according to Cooke (1924), there are few if any identical species of mollusks in the two regions. The faunas of both areas appear to be of tropical and subtropical aspect.

Thus the character and relationships of the Eocene faunas around the middle and north Atlantic do not appear to be compatible with Wegener's proposed distribution of the continental areas and seaways, although Rutsch (1940) from consideration of a few Eocene species from the Caribbean and northern South America believes that the distribution of the forms he cites is best explained by Wegener's hypothesis. However, if Rutsch is correct in ascribing the relationships with which he is concerned to "floating continents", why are there not many such instances instead of a few in faunas which most certainly must have numbered many hundreds of species if not thousands? Hertlein (1937) recognized about 29 species of mollusca and echinoids as occurring along the western shores of the Americas and adjacent islands, and in Polynesia and the Indo-Pacific. This is a far greater list of species than Rutsch discusses, and the distances over which they have had to travel are greater than those necessary for migration between West Africa and South America today, yet no one has seriously suggested that the various Pacific areas from which Hertlein's material was recorded were ever in close geographic proximity to one another. The explanation of the wide distribution of species such as those recorded by Hertlein, and those discussed by Rutsch, must lie within the realm of some other hypothesis than that of continental drift.

It would appear that the climatic and biologic relationships of the various Cenozoic faunas of the Pacific basin, together with the lack of more marked affinities between Tertiary faunas of the north Atlantic basin, strongly militate against either Wegener's, Grabau's, or DuToit's versions of "Continental Drift", insofar as these theories postulate a different distribution of the land masses and different positions of the poles during the Tertiary than at present.

## Climatic Criteria

In this paper, only criteria of an organic nature will be used. In considering and comparing past and present marine temperatures, reference should be made to Vaughan's (1940, p. 439–450) discussion of marine environments and climatic zones. Inasmuch as the primary thesis of this paper deals with past extensions and restrictions of the tropical and subtropical zones, the most important criteria are those indicating tropical or subtropical temperatures.

Foremost among the indicators of such conditions are the reef corals and related types. Their thermal distribution has been discussed by Vaughan (1940, p. 458–464), and Vaughan and Wells (1943, p. 52–63), who show that the average minimum (in February) temperatures *endurable* by a vigorous growth of reef corals is 18.5° C., and that they cannot survive this temperature for any length of time. These authors have used the term "hermatypic corals" for such forms. Most West Indian species of hermatypic corals are killed by temperatures lower than 11° C. (Vaughan and Wells, 1943, p. 55), while the maximum temperature endurable is about 36° C. At the present time, such conditions are found on the Pacific Coast only south of Cape San Lucas (Fig. 2). These authors also note (p. 52) that active building of coral reefs ceases at depths below 46 meters and that most reef-building takes place in waters less than about 28 meters deep. However, certain hermatypic corals may live down to depths of 90 meters. The genera of hermatypic corals which may live in waters of this depth include *Pocillopora*, *Pavona*, *Leptoseris*, *Montipora*, *Porites*, *Madracis*, and *Fungia*. By inference, therefore, it may be expected that the hermatypic corals which are found in the marginal areas of the tropical belts will also be members of this group, for these are corals which

can withstand somewhat lowered temperatures. Such is certainly the case in the Gulf of California, where *Porites* is found in considerable abundance in areas where the minimum temperatures are around 15° C. (Durham, 1947, p. 22); *Pavona* lives in areas of minimum temperatures of about 16° C. (Durham, 1947, p. 20); and *Pocillopora* and *Psammocora* in areas with minimum temperatures of about 17° C. (Durham, 1947, p. 16, 19). It is well to note that the reef or hermatypic corals do not necessarily live only in reefs, but rather in areas where the temperature is above a certain minimum, the depth of the water is less than the critical depth of the genus, and certain other conditions are fulfilled. Within these limits coral reefs develop under certain specific conditions. It should also be noted that no true coral reefs have yet been found in the Cenozoic deposits of the Pacific Coast north of the Gulf of California area, although there are a goodly number of typical reef coral genera recorded and the faunas of certain localities indicate a probable proximity to reefs.

Vaughan (1940, p. 458) has noted that the orbitoidal foraminifera are tropical or subtropical organisms that lived in waters of less than 180 meters depth. In the Cenozoic record of the Pacific Coast, most of the occurrences of hermatypic corals have been in strata which contain orbitoidal foraminifera, often in considerable abundance; further, these foraminifera occur only in strata which can be shown by other means to have been deposited in tropical or subtropical temperatures.

Among the mollusca, there are numerous genera[1] in the recent faunas, most of whose species are particularly characteristic of tropical (20° C. or warmer) waters. Among the more important of these are the following gastropods and pelecypods:

GASTROPODA

| | |
|---|---|
| *Ancillaria* | *Cancilla* |
| *Architectonica* | *Cassis* |
| *Astralium* | *Cerithium* |
| *Atilia* | *Chicoreus* |
| *Canarium* | *Concholepas* |
| *Costellaria* | *Phos* |
| *Distorsio* | *Phyllonotus* |
| *Fasciolaria* | *Pterocera* |
| *Ficus* | *Pustularia* |
| *Gemmula* | *Pyrene* |
| *Glyphis* | *Rimella* |
| *Gyrineum* | *Senectus* |
| *Harpa* | *Solenosteira* |
| *Latirus* | *Strioterebrum* |
| *Modulus* | *Strombus* |
| *Monodonta* | *Terebellum* |
| *Morula* | *Theliostyla* |
| *Morum* | *Tribulus* |
| *Nerita* | *Turbo* |
| *Neritina* | *Voluta* |
| *Oliva* | *Xenophora* |

PELECYPODA

| | |
|---|---|
| *Codakia* | *Miltha* |
| *Divaricella* | *Pinna* |
| *Harvella* | *Plicatula* |
| *Lyropecten* | *Spondylus* |
| *Macrocallista* | |

Among the preceding genera are rare species which may live in temperatures slightly below 20°C, but generally speaking the members are distinctly tropical in their habitats. Genera mostly tropical in occurrence, but with one or two species which habitually occur in regions where the temperature may drop as low as 12°–15°C., include, among the gastropods, the genera *Conus*, *Cymatium*, *Cypraea*, *Mitra*, *Terebra*, *Turritella*, and *Vasum*—genera which have most of their species in the tropics, and yet characteristically have one or two species ranging farther north, and, among the pelecypods, *Americardia*, *Anadara*, *Chione*, *Crassatellites*, *Dosinia*, *Labiosa*, *Laevicardium*, *Megapitaria*, *Mexicardia*, *Pedalion*, *Pitar*, *Pteria*, *Trachycardium*, and *Trigoniocardia*.

For direct comparison in the later Tertiary and Pleistocene of the Pacific Coast, the two following lists of common genera and their maximum northward range in the Recent Pacific Coast fauna have been compiled from Keen (1937):

*Northern Limit—Point Conception (34° N. lat.)*

Gastropods

| | |
|---|---|
| *Aclis* | *Cyclostrema* |
| *Anachis* | *Cymatium* |
| *Bulla* | *Engina* |

[1] In the following lists, genera and subgenera are considered as of coordinate rank.

| | |
|---|---|
| *Forreria* | *Seguenzia* |
| *Potamides* | *Terebra* |
| *Pyrene* | *Vermicularia* |

Pelecypods

| | |
|---|---|
| *Americardia* | *Chione* |
| *Amiantis* | *Crassatellites* |
| *Anadara* | *Dosinia* |
| *Anatina* | *Laevicardium* |
| *Apolymetis* | *Pedalion* |
| *Atrina* | *Pteria* |
| *Bathyarca* | *Trigoniocardia* |

*Northern Limit—Monterey Bay (37° N. lat.)*

Gastropods

| | |
|---|---|
| *Alabina* | *Liotia* |
| *Borsonella* | *Marginella* |
| *Caecum* | *Pyramidella* |
| *Cypraea* | *Simnia* |
| *Cytharella* | *Sinum* |
| *Daphnella* | *Turritella* |

Pelecypods

| | |
|---|---|
| *Barbatia* | *Tagelus* |
| *Corbula* | *Tivela* |
| *Cyathodonta* | *Trachycardium* |
| *Donax* | *Ventricola* |
| *Sanguinolaria* | |

Fossil nautiloids are generally considered to be indicators of tropical and subtropical temperatures—the living *Nautilus* is certainly confined to such regions and the faunas associated with the extinct Tertiary genera and species confirm this conclusion.

In addition to the preceding groups of invertebrates, much evidence is available in Paleobotany and occasional bits of evidence are found in other fields. Palm wood is often found fossil and is a valuable indicator of subtropical or tropical climates (*See* Bartholomew, 1942, p. 17, "Northern Limit of Palms"). The larger reptiles are usually confined to tropical or warm temperate areas and as such the record of a crocodilian (Durham, 1944, p. 196) from the lower Oligocene of Washington is in agreement with the evidence of the associated fauna in indicating at least a subtropical climate. Fossil bird assemblages are useful climatic indicators (Miller, 1940), and in fact the remains of almost any organism which was not uryopic are significant.

### Limitations

In making any interpretation of past marine climates, one is obviously limited by the extent of fossiliferous sediments of the epoch being considered. The general distribution of Tertiary marine sediments of the Pacific Coast is indicated on Figure 2. In general, marine sediments are unknown north of the Canadian boundary (49° North latitude), except for a small area at the head of the Gulf of Alaska and isolated areas in the Aleutian Islands (not shown on map). The small area at the head of the Gulf of Alaska extends from about 58°30′ to about 60° N. latitude. The best known Tertiary deposits of the region are of upper Oligocene age (Clark, 1932), but Pliocene, possibly middle (called Miocene, but Pliocene in present-day terminology) fossiliferous strata have been reported (Mertie, 1931). On the basis of unpublished reports, it appears that lower Oligocene strata also occur in that vicinity but their fauna is unknown. Pleistocene marine deposits occur irregularly at least as far north as Juneau, Alaska (Dall, 1904), while (Dall (1904) reports Eocene and Miocene fossils from the Alaskan Peninsula area and the Shumagin Islands in the vicinity of 56° N. latitude, and 160° W. longitude. Although fossiliferous strata representing various parts of the Tertiary sequence are known to occur south of 32° N. latitude, their faunas are in general not well known.

With respect to the smaller divisions of Cenozoic time, except for the Pleistocene, the "stages or formations" of column 1 on the left side of the chart prepared by Weaver, *et al.* (1944) have been used as a standard in as far as possible. Within the Pleistocene, the various methods of correlation now in use are not refined enough to permit a reasonable synthesis of the available data.

Fossiliferous sediments representing the Martinez and Meganos stages of the Paleocene are known only as far north as latitude 40°, imposing serious limitations on interpretations of their faunas. Within the Eocene and Oligocene, all of the stages are represented as far north as 49° latitude, thus affording a much

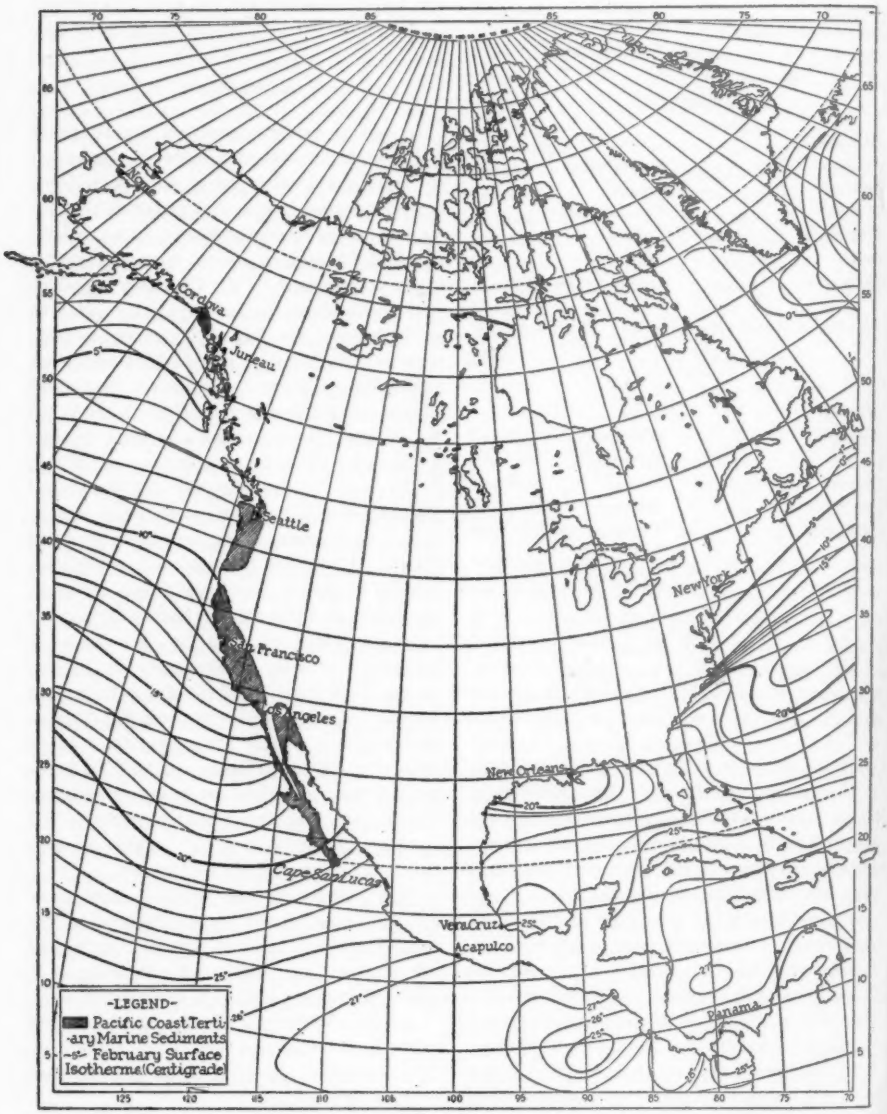


FIGURE 2.—DISTRIBUTION OF TERTIARY MARINE SEDIMENTS ON PACIFIC COAST AND FEBRUARY SURFACE ISOTHERMS IN WATERS ADJACENT TO NORTH AMERICA

firmer basis for deductions, particularly with the representation of the upper Oligocene at the head of the Gulf of Alaska. The oldest Miocene (Vaqueros) has an extremely limited northward distribution, with the fauna being known only to about 37° 30′ N. latitude. However, some data (Durham, Harper, and Wilder, 1942) indicates the possibility of its occurrence at about 45° N. latitude. The succeeding middle Miocene to middle Pliocene are all found

intermittently along the Pacific Coast to 49° N. latitude, with the middle Pliocene also being represented near the head of the Gulf of Alaska (Mertie, 1931, as "upper Miocene"), and middle (?) Miocene in the Queen Charlotte Islands (Dawson, 1880). Definitely upper Pliocene beds are known in the vicinity of San Francisco (38° N. lat.), and farther south in California, and a late Pliocene marine fauna is known (unpublished data) from near 41° N. latitude on the California Coast. The Quinault and Quillayute faunas of Washington (48° N. latitude) may be in part upper Pliocene also. Pleistocene marine deposits are variously represented at least as far north as Juneau, Alaska (Grant and Gale, 1931; Dall, 1904).

From the foregoing discussion, it can be seen that the marine sediments of the Pacific Coast are more or less widely distributed for the various epochs of time to be considered, and that the interpretations for those same divisions will be accordingly of varying degrees of preciseness. The evidence available of course is similarly controlled by the extent to which the faunas have been studied in the different geographic parts of their ranges. Generally speaking, the faunas from the southern half of California have been most extensively studied, but nowhere along the Pacific Coast, with the possible exception of the Pleistocene of the Los Angeles area, have they been studied to an extent comparable with the well known European faunas of the same epochs. It is highly probable that the marine faunal picture of the Pacific Coast Tertiary is less than 50 per cent complete, and most likely not more than 25 per cent complete. The dominant elements of the various faunas are fairly well known however—it is the less common and smaller members that remain to be recorded.

## Climatic Measurement

In considering the past marine climates of the Pacific Coast, the position of the 18° C. isotherm in February has been taken as the most easily determinable point during the early Tertiary because of its close approximation to the minimum temperatures endurable by reef or hermatypic corals, which in turn are probably the most sensitive (thermally) of the organisms available for consideration. The position during the month of February is used, as this is the coldest period of the year and the time at which the isotherms, on the average, are farthest south. Conversely, during the month of August, the isotherms are farthest north. For details of this northward shift in summertime, charts II and III in Sverdrup, Johnson, and Fleming (1942) should be consulted. At the present time the February 18° C. isotherm falls near Cedros Island (= 28° N. lat.) on the coast of Lower California, some 450 miles south of Los Angeles. This point corresponds to the southern limit of the "Southern California" fauna of Smith (1919, p. 133–134), but falls about in the middle of Bartsch's (1912, p. 299) Californic faunal area, and is within Dall's (1921, p. 1) Tropic "division". In Dall's terminology, this point might presumably be the southern boundary of the subtropical, as distinguished from purely tropical faunas. Actually the biggest faunal breaks along this part of the Pacific Coast fall either at Cape San Lucas (Bartsch, 1912, p. 299, boundary between Californic and Mazatlanic faunal areas; Durham, 1947, p. 9) or at Point Conception (Bartsch, 1912, p. 299; Dall, 1921, p. 1). These two points correspond respectively to the positions of the February 20° C. and 12.5° C. isotherms. The Recent molluscan genera whose northward ranges end at Point Conception are listed on an earlier page.

Using the criteria previously discussed, the following interpretations are presented:

## Mesozoic Climate

As background for the Tertiary, a few comments on the climatic implications of the Pacific Coast Mesozoic faunas are presented here, although the inferences are on a much more uncertain footing than in later epochs where the organisms are more closely related to living faunas. Smith (1927, p. 7) makes the following statement with regard to the lower Noric (uppermost Triassic) coral zone:

"In all these regions [Mediterranean, California, Alaska, Himalayas] it is characterized by the abundance of reef-forming Astraeidae, nearly related to modern reef-building forms that do not flourish in water that has a temperature below

74° F. It is highly probable that this equable temperature extended to at least 61° North Latitude in the North Pacific."

Data on the Jurassic faunas is still poor and the interpretations offered by Smith (1910) are still the most comprehensive. He notes (p. 483) that there are reef corals in the middle Jurassic of the Queen Charlotte Islands (53° N. lat.), and that in the upper Jurassic and lower Cretaceous the widespread *Aucella* faunas are of a boreal type (Pavlow, 1907, p. 84). At the same time, Smith notes the inconsistency of cycads, which presumably required warm climates, occurring in the same strata as the aucellan fauna. Skepticism regarding the boreal implications of this fauna (*Aucella*) has also been expressed by others workers, but Anderson (1938, p. 73–75) still accepts this interpretation. The first post-aucellan fauna of California, the Horsetown, is considered by Anderson (1938) to be of "southern or subtropical aspect".

Within the upper Cretaceous, there are several indications that the climate was either subtropical or tropical fairly far to the north. Whiteaves (1900, p. 305–307) has recorded a typical reef coral, *Astrocoenia irregularis* Whiteaves, from the Queen Charlotte Islands (53° N. lat.), a nautiloid (*Cymatoceras*), numerous ammonites, the pelecypod genera *Corbula, Nemodon, Trigonarca, Cucullaea, Melina, Gervillia, Meleagrina* and *Avicula*. The general assemblage, even though there may be some doubt regarding the association of everything listed, appears to indicate a subtropical environment. In addition, Whiteaves (1903, p. 407–409) has listed from the southern part of Vancouver Island and the Sucia Islands (49° N. lat.) the following: the nautiloid *Cymatoceras;* numerous ammonites, the gastropods *Cypraea, Cerithium, Potamides;* the pelecypods *Corbula, Laevicardium, Crassatellites,* "*Arca*", *Cucullaea, Pinna, Perna, Meleagrina,* and *Spondylus*. The assemblage appears adequate to indicate a temperature of 20° C.

In California, F. M. Anderson has recorded at least 3 species of the gastropod genus *Cypraea* in the upper Cretaceous (Ingram, 1947, p. 61, footnote). The reef coral genus *Favites*, a number of species of nautiloids, and numerous ammonites are present, and in addition to the preceding, there are large pelecypods and gastropods that are probably indicative of tropical or subtropical conditions in the various upper Cretaceous faunas. In general, it appears probable that the 18.5° C. isotherm was some place in the vicinity of 53° N. latitude, or even farther to the north during this epoch. This is in general accord with the temperatures indicated for the upper Cretaceous seas of England by Urey (1948, p. 495).

## Tertiary Climate

### *Paleocene*

*Martinez.*—As indicated previously, fossiliferous strata of this age are not known north of 40° latitude, and even at this point the fauna is not well known. The northernmost fauna which is at all adequately known is at Lower Lake (approximately 39° N. lat.), California. This fauna was described by Dickerson (1914), who lists a possible reef coral (*Stylophora?*). This coral has been re-examined and while it is apparent that it is not a member of the genus to which he referred it, it is equally certain that it is a colonial hermatypic coral. As other members of the fauna, he lists the solitary coral, *Flabellum remondianum* Gabb; the pelecypod genera *Crassatellites, Perna, Pinna, Plicatula,* and *Venericardia;* the gastropod genera *Cypraea, Turritella,* and *Tritonium*. The total assemblage present appears to be characteristic of temperatures of 20° C. or slightly higher—that is, the 18.5° C. isotherm must have been some distance to the north of this latitude. This minimum temperature is corroborated by the occurrence of the reef coral genus *Haimesiastraea* in the vicinity of the town of Martinez (lat. 38°), and the ecologically similar genera *Goniopora* in the Panoche Pass area (36°30′ N. lat., unpublished data) and *Siderastrea* in the Simi Valley (Faustino, 1931).

In addition to the corals, Ingram (1947) notes that three species of the gastropod genus *Cypraea* have been described from the Martinez strata of Simi Valley (about 34°15′ N. lat.), while in the San Francisco Bay region (38° N. lat.), Dickerson (1914, p. 106–111) lists the pelecypod genus *Crassatellites*, the gastropod genera *Architectonica, Ficopsis, Nerita, Ovula, Turritella, Tritonium, Xenophora,* and the nautiloids *Hercoglossa* and *Eutrephoceras* (*Naut-*

*ilus* of Dickerson). The occurrence of the preceding genera in the faunas appears to amply substantiate the view that the faunas are of tropical aspect.

*Meganos.*—The only well described (Clark and Woodford, 1927) fauna of this age is found in the vicinity of Mount Diablo (just south of 38° N. latitude), although strata of this age are known from Round Valley (40° N. latitude), California. The described fauna is limited, but includes species of the pelecypod genera *Crassatellites*, *Macrocallista*, *Pedalion*, *Pitar*, *Pteria*, *Venericardia* (large); the gastropod genera *Architectonica*, *Cypraea*, *Ficopsis*, *Oliva*, *Pseudoliva*, *Turritella;* the corals *Flabellum*, *Archohelia*, and *Turbinolia*. In all, the fauna contains sufficient elements to indicate that it lived within the confines of the tropics and that the 18.5° C. isotherm must have been considerably farther north.

### *Eocene*

*Capay.*—In contrast to the preceding Paleocene, strata of Capay age and their contained faunas are known as far north as the southern end of Vancouver Island (49° N. latitude). Examination of the faunas known from this general region immediately shows their tropical character. The reef coral genera *Astreopora*, *Colpophyllia*, *Leptophyllastrea*, *Madracis*, *Montipora*, and a generically indeterminate "agaricid", along with orbitoidal foraminifera, mollusca, brachiopods, and calcareous algae are found (Durham, 1942a, p. 85) along the shore south of the straits of Juan de Fuca (slightly north of 48° latitude). The assemblage is clearly tropical in character and indicates that the 18.5° C. isotherm must have been some distance north of this region. The pelecypods *Pitar*, *Macrocallista*, *Spondylus* and the gastropods *Nerita*, *Turritella*, and *Eocernina* were found in the same fauna.

Further south, west of the town of Salem, Oregon (45° N. latitude), Capay fossils, including the reef corals *Leptoseris* and *Leptophyllastrea;* the pelecypods *Macrocallista* (?), *Pitar*, *Plicatula* (?), *Spondylus*, *Venricardia* (large); the gastropods *Nerita*, *Spiroglyphus* (?), and *Turritella;* orbitoidal foraminifera; and several species of brachiopods, have been recorded (Baldwin, 1947, p. 10–12). This fauna strongly substantiates the tropical character of the Washington material. All known faunas of this age in the more southern areas of Oregon and California are in agreement with this interpretation.

*Domengine.*—The faunas of this age in Washington are less definitive than those of Capay age, but the indicators present point to a similar environment. The coral *Madracis*, although not strictly a reef dweller, is usually restricted to tropical regions and has been recorded (Durham, 1942a) from the same area in Washington as the Capay fauna previously discussed (48° N. lat.). In addition, the pelecypod genera *Corbula*, *Macrocallista*, *Pitar*, and large *Venericardia;* the gastropod genera *Eocernina*, *Eratopsis*, *Ficopsis*, *Harpa*, *Nerita*, and abundant *Turritella;* and the cephalopod *Aturia* are known from the same beds. The total assemblage indicates much the same temperature as during Capay time.

Farther south, in central western Oregon (43°–44° N. lat.), the faunas of Domengine age listed by Turner (1938) include the following: pelecypods—large *Venericardia*, *Corbula*, *Crassatellites;* gastropods—*Pseudoliva*, *Ficopsis*, *Cypraea*, *Turritella*, *Conus*, *Eocernina*, *Rimella*, *Ancilla*, and *Architectonica*. In California, the reef corals *Astreopora*, *Astrocoenia*, *Coeloria*, *Favia*, *Leptastrea*, *Oulophyllia*, *Podasteria*, and *Stylophora* have been recorded from Kern County (35° N. lat.) (Durham, 1942b). This assemblage is decidedly tropical in character and probably represents a minimum temperature above 25° C. The molluscan faunas listed by Vokes (1939) from the type area of the Domengine in the vicinity of Coalinga (36°–37° N. lat.) confirm this conclusion, containing the following: pelecypoda—*Pinna*, *Pteria*, *Pedalion*, *Spondylus*, *Crassatellites*, large *Venericardia*, *Miltha*, *Macrocallista*, *Pitar*, *Pelecyora*, *Corbula;* gastropods—*Akera*, *Terebra*, *Conus*, *Olivella*, *Ancilla*, *Voluta* ?, *Lyria*, *Harpa*, *Pseudoliva*, *Ficopsis*, *Cypraea*, *Rimella*, *Terebellum*, *Potamides*, *Turritella*, *Architectonica*, *Xenophora*, *Eocernina*, *Nerita*, *and Velates*. In all, the assemblage listed by Vokes is typically tropical in character and the minimum temperature here suggested is in general agreement with Vokes'. conclusions (1940, p. 603). By

analogy with the distribution of the present day isotherms where there is about 4° C. difference in temperature between the latitude of the type Domengine and the latitude (48° N.) of the northern Washington outcrops, it may be reasonably inferred that the temperature in the northern area during Domengine time was about 21° C., thus substantiating the conclusion based on the Washington faunas that the 18.5° C. isotherm was some distance farther north.

*"Transition".*—The faunas of this "stage" have not been segregated as yet, but the known elements do not differ greatly from the members of the preceding Domengine and following Tejon faunas, indicating that similar temperatures must have prevailed.

*Tejon.*—This stage is represented in Washington and northwestern Oregon by the Cowlitz formation and its equivalents, the faunas of which have much the same aspect as those of the Domengine stage, except that they are generally of somewhat deeper-water character and in consequence usually lack the shallow-water corals and molluscs. However a single reef coral, *Astreopora*, has been recorded (Durham, 1942a, p. 86) from near Seattle (47°30′ N. lat.), and appears to indicate that the 18.5° C. isotherm was still north of this vicinity. Float boulders containing the pelecypod *Crassatellites*, and other fossils of Cowlitz age have been found near the northeastern corner of the Olympic mountains (48° N. lat., unpublished data). The type area (46°30′ N. latitude) of the Cowlitz formation furnishes many well preserved fossils (Weaver, 1943) which includes the following: pelecypoda—*Obliquarca*, *Pteria*, *Crassatellites*, large *Venericardia*, *Macrocallista*, *Pitar*, *Sanguinolaria*, *Corbula;* gastropoda—*Nerita*, *Neritina*, *Turritella*, *Potamides*, *Ficopsis*, *Cymatium*, *Cantharus*, *Pseudoliva*, *Conus;* nautiloids—*Cimomia* (unpublished data), *Aturia*. Near Balch, Washington, the hermatypic coral genus *Hydnophora* has been collected (unpublished data) from the lower part of the Cowlitz formation. The foregoing assemblages indicate temperatures of 20° C. or slightly higher, and corroborates the occurrence of *Astrocoenia* and *Crassatellites* slightly farther north. The known faunas of Cowlitz age in Oregon have the same composition as those of Washington and add no new data.

In California, the best known faunas of this age are those of the type Tejon (approximately 35° N. lat.) at the southern end of the San Joaquin valley, described by Anderson and Hanna (1925). The fauna includes the following: pelecypoda—*Corbula*, *Crassatellites*, *Divaricella*, *Macrocallista*, *Pitar*, *Pteria*, *Spondylus*, large *Venericardia;* gastropoda—*Architectonica*, *Conus*, *Rimella*, *Cypraea*, *Ficopsis*, *Ficus*, *Pseudoliva*, *Terebra*, *Turritella*, *Voluta* ?; nautiloid—*Aturia*. Marks (1941) has also reported the nautiloid *Cimomia* and the large foraminiferan *Discocyclina* from the same section. Although less definitive in its character than the Domengine faunas of the San Joaquin Valley, this fauna is clearly tropical in its aspect and as such is in harmony with the character of the more northern faunas.

*Keasey.*—The faunas of this stage and their distribution are poorly known as yet, although it is known that they are best developed in northwestern Oregon and that they are generally similar to those of the preceding Cowlitz of that area, except that much of the Keasey formation was deposited in deeper water (for example: Durham, 1942a, p. 87, *Keasey shale*). The faunas (Durham, 1937, 1942a; 1944; unpublished data) of the type Keasey (46° N. latitude) includes the following: foraminifera—*Operculina;* corals—*Stephanocyathus*, *Flabellum* (of *F. pavoninum* group), *Astrangia;* pelecypoda—*Pitar*, *Crassatellites;* gastropoda—*Conus*, *Turritella;* nautiloids—*Aturia*, *Eutrephoceras*. The preceding list is not large, but is sufficient to indicate that the waters in which the fauna lived were above 18.5° C.

Faunas of the same age are equally poorly known in California. Undescribed mollusca in the University of California collections from the type locality of the Gaviota formation (about 34°30′ N. lat.) include the following: pelecypoda—*Crassatellites*, *Macrocallista*, *Pitar*, *Tivela*, large *Venericardia;* gastropoda—*Ficopsis*, *"Potamides"*, *Rimella*, *Turritella*. This list, although small, is definitely tropical in character. From the known data, it appears that the isotherms during Keasey time must have had a position intermediate between those of preceding Tejon time and following Lincoln

time, therefore, the 18.5° C. isotherm must have been at 48° North latitude or farther north.

### *Oligocene*

*Lincoln.*—The faunas of this and the following stage (Blakeley) are best known in Oregon and Washington, although they do occur in limited areas of California. In Washington the fauna of the Quimper sandstone (Durham, 1944, p. 117–120) at 48° N. latitude includes the following: corals—*Siderastrea ?*, *Archohelia*, *Tubastrea*, *Astrangia*, indeterminate colonial eupsammid genus; pelecypoda—*Arca*, *Corbula*, *Pitar*, *Sanguinolaria;* gastropoda—*Conus*, *Cypraeogemmula*, *Gemmula*, *Marginella*, *Ficus*, *Nerita*, *Phalium*, *Seraphs*, *Turritella;* nautiloid—*Aturia;* vertebrata—*crocodilian* sp. (Durham, 1944, p. 196). The assemblage indicates a minimum temperature of about 20° C. This is further corroborated by the occurrence of fossil palm wood in this vicinity (Durham, 1942a, p. 88).

Further south (46°30′ N. latitude) in Washington, the Gries Ranch fauna described by Effinger (1938) includes the following: pelecypoda—*Chama*, *Corbula*, *Crassatellites*, *Macrocallista*, *Pedalion*, *Pitar*, *Semele;* gastropoda—*Conus*, *Cypraeogemmula*, *Marginella*, *Phalium*, *Terebellum*, *Volvulella*. This assemblage confirms the tropical character of the Quimper sandstone fauna. It is unfortunate that although unpublished data indicates the occurrence of faunas of this age at the head of the Gulf of Alaska, their character is unknown. Weaver (1943, p. 638–650) has also recorded the gastropods *Architectonica*, *Cypraea* and *Persicula* from strata of this age in Washington. It appears that at this time the 18.5° C. isotherm must still have been north of 49° North latitude.

### *Upper Oligocene—Lower Miocene*

*Blakeley.*—The best known faunas of Blakeley age occur in the area from Seattle, Washington, to the southern end of Vancouver Island (48°–49° N. latitude). The faunas of strata referred to this stage have been described by Clark and Arnold (1923); Clark (1925); Tegland (1933); and Durham (1944). The reef coral genus *Siderastrea* has been recorded several times (Durham, 1942a, p. 88–89), besides the ecologically similar genus *Eusmilia*. In addition the recorded fauna includes the following: pelecypoda—*Antigona*, *Chione*, *Corbis*, *Crassatellites*, *Macrocallista*, *Pitar*, *Semele*, *Tivela;* gastropoda—*Acteon*, *Ancilla*, *Ficus*, *Marginella*, *Oxystele*, *Rapana*, nautiloid—*Aturia*. Using the criteria presented in this paper, the Blakeley fauna of this area appears to represent a minimum temperature of not less than 18.5° C.

Faunas of this age have also been reported from the head of the Gulf of Alaska (60° N. latitude) by Clark (1932), and include the following: pelecypoda—*Chione*, *Macrocallista*, *Pitar;* gastropoda—*Turritella;* nautiloid—*Aturia*. This list of genera, while small, still seems to indicate warm temperatures, probably a minimum of not less than 15° C. This is particularly notable in view of the fact that Taliaferro (Clark, 1932, p. 799) concluded that at least part of the containing sediments were apparently of glacial origin. It would appear that the climatic conditions prevailing during the upper Oligocene in the head of the Gulf of Alaska must have been similar to those of the present epoch in the South Island of New Zealand where there are extant glaciers and yet the minimum yearly temperature is about 14° C. (Sverdrup, Johnson, and Fleming, 1942, Chart 2). It does not appear that Clark's comparison (1932) of the fossil fauna with that living off the area at the present time is justified, for he apparently neglected to consider the genera listed above. Another possible explanation which could account for the temperatures indicated for the Gulf of Alaska region may be found in the distribution of the August isotherms along the Pacific Coast at the present time (Sverdrup, Johnson, and Fleming, 1942, Chart 3). At this time of the year, the 15° C. isotherm leaves the California coast at about 34° N. latitude, goes out to sea, and then swings in close to the coast again in latitudes 44° to 48°. It may be that with the more northward position of the isotherms during upper Oligocene times a similar isothermal pattern was in effect, but farther north along the coast.

*Miocene*

*Vaqueros.*—As indicated previously, typical faunas of this age are not recognized north of about 37°30′ N. latitude. The monograph by Loel and Corey (1932) records the following critical members of the fauna: pelecypoda—*Amiantis, Antigona, Anadara, Trachycardium, Chione, Clementia, Corbis, Crassatellites, Dosinia, Apolymetis, Lyropecten, Pedalion, Miltha, Pinna, Pitar, Pteria, Sanguinolaria, Spondylus, Tivela;* gastropoda—*Architectonica, Cerithium, Conus, Cypraea, Ficus, Trophosycon, Nerita, Oliva, Potamides, Rapana, Solenosteira, Terebra, Turritella, Voluta, Xenophora;* coral—*Oculina.* The fauna as a whole clearly indicates minimum temperatures of not less than 20° C., and enough of the genera are found in the northernmost outcrops to indicate that the aforesaid isotherm must have been at least that far (37°30′ N. lat.) north. By comparison with the distribution of the present-day isotherms, it appears that the 18.5° C. isotherm must have been at about 40° N. latitude. One of the notable characters of the Vaqueros fauna is the absence of hermatypic corals, or indeed of corals of any type, for only a single genus (*Oculina*) has been reported, despite intensive collecting. The environment, in some phase, perhaps salinity, perhaps turbidity, was apparently not favorable to coral growth.

*Temblor-Briones.*—In contrast to the preceding Vaqueros, marine sediments (Clallam formation) of Temblor age are found as far as the Straits of Juan de Fuca (48°15′ N. latitude). The fauna recorded (Weaver, 1943; Durham, 1944, p. 129) from this area include the following: pelecypoda—*Anadara, Chione;* gastropoda—*Turritella, Trophosycon;* nautiloid—*Aturia.* Farther south, in southwestern Washington and northwestern Oregon (45°–47° N. lat.), the Astoria formation has the following fauna: coral—*Lophelia* (unpublished data); pelecypoda—*Anadara, Chione, Pitar;* gastropoda—*Ficus, Turritella, Trophosycon, Olivella, Acteon;* nautiloid—*Aturia.* On the basis primarily of the coral, *Anadara, Chione, Ficus,* and *Turritella,* it would appear that the 13° C. isotherm must have been at about 48° N. latitude.

The "Astoria Miocene" reported by Dall (1896) from various localities in the Alaskan Peninsula, the Aleutian Islands, the Yukon Valley, Kodiak Island, Lituya Bay, and Queen Charlotte Islands, has never been adequately studied or reported upon. The Lituya Bay and Kodiak Island occurrences have had more modern faunal lists (Mertie, 1931; Capps, 1937) reported, and as noted further on might be Pliocene instead of Miocene in age. Dawson (1880, p. 87) lists a considerably larger fauna than Dall, from the Queen Charlotte Islands, and on the basis of "*Arca*" *microdonta* the fauna would appear to be of Miocene age rather than Pliocene. This species would be referred to the genus *Anadara* at the present time, and very possibly is a synonym of one of the other more common Pacific Coast Tertiary species. A Miocene age rather than a Pliocene one appears more probable because the genus *Anadara* is not known north of central Oregon during the Pliocene, but ranges up at least as far north as 48° N. latitude (Clallam formation) during the middle Miocene. The Queen Charlotte Islands occurrences are between 53° and 54° N. latitude, and on the basis of the *Anadara* would appear to have been laid down in temperatures of not less than 12° C. It appears probable that Dall's "*Astoria Miocene*" includes strata of several different ages.

Farther south, in Kern County, California (approximately 34°30′ N. lat.), the faunas of this age (Loel and Corey, 1932; Keen, 1943) include the following: pelecypoda—*Anadara, Chione, Dosinia, Miltha, Atrina, Chama, Clementia, Crassatellites, Donax, Apolymetis, Lyropecten, Pitar, Semele;* gastropoda—*Cerithium, Conus, Ficus, Trophosycon, Melongena, Oliva, Terebra, Turritella;* nautiloid—*Aturia.* In the San Francisco Bay area (38° N. lat.), the Briones formation (Trask, 1922) has the following fauna: pelecypoda—*Antigona, Anadara, Apolymetis, Chione, Dosinia, Lyropecten, Pitar, Tivela;* gastropoda—*Acteon, Trophosycon, Oliva.* It would appear that the Kern County faunas lived in a minimum temperature of 19° or 20° C., and that the Briones fauna reflects a temperature of not less than 17° or 18°, thus in general corroborating the interpretation of the Washington faunas.

*Cierbo-Neroly.*—Although the faunas of these

two epochs may be separated in California on the basis of different species of echinoids, pectens, and certain other molluscs, they probably represent small units of time and are here considered together. In Washington, the lower part of the Montesano formation (47° N. latitude) is apparently equivalent to the Neroly. Its fauna includes (Weaver, 1943) the following: pelecypoda—*Chione, Semele;* gastropoda—*Phalium, Nassarius, Olivella, Megasurcula.* The *Phalium* and *Megasurcula* might indicate an environment about the same as that off San Francisco today (11° C.).

The Cierbo-Neroly fauna (38° N. lat.) was described by Clark (1915), and includes the following: pelecypoda—*Amiantis, Andara, Apolymetis, Atrina, Chione, Dosinia, Lyropecten, Pitar, Sanguinolaria, Tivela;* gastropoda—*Acteon, Astraea, Bursa, Trophosycon, Forreria, Turritella.* This assemblage appears to indicate a minimum temperature of about 13° C., a conclusion essentially in harmony with that expressed by Clark (p. 427–428).

### *Pliocene*

*Jacalitos.*—It appears probable that the upper Montesano formation in Washington, and the lower Empire formation (43°30′ N. latitude) in Oregon, are about equivalent in age to the Jacalitos formation of Central California (36° N. latitude). In Alaska, the beds reported by Mertie (1931) from the head of the Gulf of Alaska, and those reported by Capps (1937) from Kodiak Island, may be in part of the same age. The fauna of the Empire formation (Howe, 1922) contains the following: pelecypoda—*Anadara, Chione;* gastropoda—*Astraea, Cymatium, Gyrineum, Olivella, Phalium, Turcica, Sinum.* This assemblage might indicate a minimum temperature of 12° for the fauna of the Empire formation.

The fauna of the Jacalitos formation (Nomland, 1916b, 1917b) includes the following: pelecypoda—*Anadara, Chione, Dosinia, Apolymetis, Lyropecten, Sanguinolaria, Semele, Tivela;* gastropoda—*Astraea, Forreria, Megasurcula, Trophosycon, Sinum, Turritella.* The assemblage would appear to indicate a minimum temperature of not less than 13°C.

*Etchegoin.*—The type area for strata of this age is around Coalinga (36° N. latitude), but they are also represented north of San Francisco Bay (Wilson's Ranch, 38°30′ N. latitude), and in the Wildcat formation of northwestern California (41°–42° N. latitude). The Empire formation of Oregon also appears to be in part of this age—its fauna has been discussed under the Jacalitos. The fauna of the Wildcat (B. Martin, 1916) includes the pelecypods *Anadara* and *Chione,* which would indicate a minimum temperature of about 12° C.

The Etchegoin of the Coalinga-Kettleman Hills region (36° N. latitude) has the following critical fossils in its fauna (Woodring, Stewart, and Richards, 1940): pelecypoda—*Arca, Anadara, Lyropecten, Apolymetis, Trachycardium, Sanguinolaria;* gastropoda—*Forreria, Turritella.* This assemblage suggests a minimum temperature of 13° or more probably 14° C., when one considers the presence of *Lyropecten,* and the fact that the *Apolymetis* is, by inference (Woodring, Stewart, and Richards, 1940, p. 93), more closely related to a southern species of *Apolymetis,* than it is to the one living off of the Los Angeles area.

*San Joaquin*—The type area of this interval is also in the vicinity of Coalinga and the Kettleman Hills (36° N. latitude). The fauna (Woodring, Stewart, and Richards, 1940, Checklist, p. 38) include the following: pelecypoda—*Anadara, Pecten* s.s. (related to a Gulf of California species), *Corbula, Semele* (*S. flavascens*?) *Trachycardium;* gastropoda—*Turcica, Murex* (cf. *M. festivus*), *Forreria;* corals—*Astrangia, Rhizopsammia* (Durham, 1941). The assemblage is adequate to indicate a minimum temperature of about 13° C.

The "*Mya japonica*" zone of the Kettleman Hills-Coalinga region has often been considered as a "cold water" interval (notably Grant and Gale, 1931, p. 53, 60, 61), but on the premises used in this paper, that conclusion is not warranted. The "*Mya japonica*" zone is so called because of the great abundance of this pelecypod in certain intervals and the scarcity of other fossils. In actuality it is not a single zone, but many *Mya* bearing beds occur scattered throughout both the Etchegoin and San Joaquin (Woodring, Stewart, and Richards, 1940, p. 29) and alternate with beds carrying a fauna of warm water character (on the basis

of the criteria here used). Furthermore the fauna of the upper *Mya* zone (as the term is used by Woodring, Stewart, and Richards) includes the genus *Anadara* (Woodring, Stewart and Richards, 1940, p. 30), while lower in the section the same *Mya*, although in lesser numbers, occurs in association with all members of the assemblage listed in the preceding paragraph. It appears that the *Mya* zone represents a brackish facies as suggested by Woodring, Stewart, and Richards (1940, p. 30, 100–101) and not a "cool-water interval".

With respect to the lower Merced fauna, considered by Smith (1919) and others, as representing cooler conditions than those which now prevail off the same area, it should be noted that the fauna (B. Martin, 1916) contains the pelecypods *Anadara trilineata*, *Chione succincta*, and *Spisula hemphilli*. On the premises used in this paper, this assemblage would indicate a temperature of not less than 12°C.

An undescribed fauna from the California coast north of Arcata (41° N. lat.) is apparently of upper Pliocene age, for it contains the extinct echinoid genus *Anorthoscutum* and the pelecypod *Arca sisquocensis* Reinhart. This fauna is notable for the occurrence in it of the *Arca*, which is related to members of the Panamic fauna, and of the coral *Dendrophyllia oldroydi* Faustino, which is living off the Los Angeles area. The aforesaid fossils would appear to indicate a minimum temperature of not less than 13°.

## Quaternary Climate

### *Pleistocene*

The data regarding the Pleistocene is susceptible to several different interpretations, depending upon the correlation of various strata and localities, and upon the interpretation of certain lines of evidence. The San Pedro Pleistocene section is the most completely known and described section on the Pacific Coast, but the varying interpretations presented confuse the picture, especially when the latest workers (Woodring, Bramlette, and Kew, 1946) do not come to any definite conclusions, and render utilization of the data from that area difficult. Individual bits of evidence from less complete sections elsewhere make it apparent, however, that there were northward and southward shifts of the isotherms which presumably correspond to glacial and interglacial epochs. Miller (1940, p. 808) indicates that the Pleistocene avifauna of Carpinteria corresponds to that existing today at Monterey. This would indicate a minimum marine temperature of about 12° C. instead of the approximately 13° C. which prevails there today. The pelecypod genus *Chione* lived in the Tomales Bay (38° N. latitude) area (Grant and Gale, 1931, p. 322) and would therefore presumably indicate a 13° C. minimum temperature instead of the 11° C. of the present time. A northward shift of the coral fauna within the Gulf of California of some 2° of latitude has been recorded (Durham, 1947, p. 11).

The coral *"Astrangia" boreas* Nomland (1917a) from the Pleistocene near Juneau, Alaska (58° N. latitude), is closely related to an undescribed living species which so far has been found only in the Channel Islands (33° N. latitude).[2] The northern occurrence of the fossil species is probably largely due to a change in position of the isotherms, although whether or not it represents a change in position of 25° of latitude is uncertain, inasmuch as there is no corroborating evidence in the small associated fauna. The genus *Dendrosmilia* is known elsewhere only by a single species from the Eocene of Western Europe, where it is a member of a fauna that is at least subtropical in aspect.

The discontinuous distribution (Mortensen, 1948, p. 355–356) of the echinoid *Echinarachnius parma* appears to be best explained on the basis of a late Cenozoic northward shift of the isotherms also. At the present time this species is found along the shores of the northern portions of both eastern and western North America, but is not found along the Arctic coasts. This distribution is best explained on the assumption that, in the not too distant past, this species had a continuous distribution along the Arctic Coast from one region to the other, presumably when the region was warmer, and that since that time the isotherms have

[2] Both species should be referred to the genus *Dendrosmilia*.

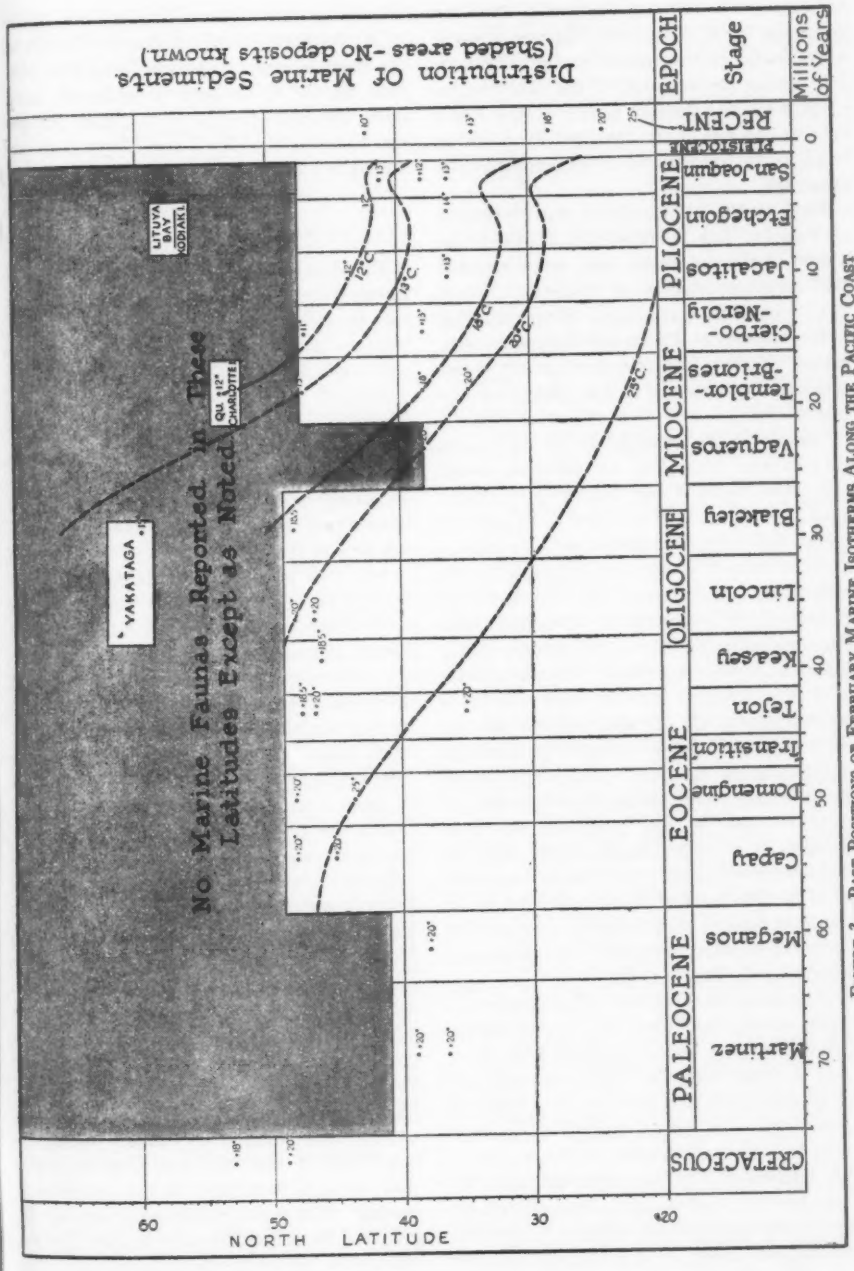


FIGURE 3.—PAST POSITIONS OF FEBRUARY MARINE ISOTHERMS ALONG THE PACIFIC COAST

migrated southward and the species could no longer live in the colder areas. This case is somewhat similar to that noted by Dall (1921, p. 5–6) among the mollusca in the northern regions, wherein a number of species now living only on one side of the continent are found as fossils in the late Pliocene or Pleistocene on the other side.

Bruff (1946) has correlated the Pleistocene of Newport Bay, California, with the Palos Verdes of the San Pedro area, and considered that its fauna indicated a temperature about 2° C. warmer than at present. Numerous other interpretations of Pleistocene faunas as indicating different temperature than at present could be cited, but the lack of precise correlation methods renders their utilization difficult.

The preceding conclusions as to marine temperatures prevailing at different places along the Pacific Coast during the Tertiary have been summarized in Figure 3. In preparation of this chart, the number of years elapsed for the different epochs, as indicated by the scale along the bottom, was determined by comparing the Weaver, *et al.* (1944) chart with the time scale presented in Simpson's (1947) paper. It is highly probable that the precise number of years assigned to each interval is not correct, but the overall total appears to be fairly reasonable. It is to be noted tha the shape of the curve for the different isotherms plotted on the chart would change if the number of years assigned to any individual epoch is changed. In consequence, the precise shape of the curve is not significant although the general trend is. The position of the isotherms in the "Recent" column on the right is their position during February, as taken from Sverdrup, Johnson, and Fleming (1942).

The position of the 25° and 20° C. isothermal curves during the Paleocene is problematical except insofar as they are known to be north of 41° latitude. They can either have remained in approximately the same position from the Cretaceous on to the middle Eocene, or there could have been a somewhat cooler interval in the Paleocene. The available data for the Paleocene does not suggest a cooling, but because of the lack of faunas north of 41° latitude, the northern limits of the tropical elements cannot be evaluated with certainty. During the lower and middle Eocene, the position of the 25° C. isotherm is fairly well controlled. From the upper Eocene on through the middle Miocene, the 20° C. isotherm is reasonably accurately determined, while in the upper Miocene and Pliocene, the 12° C. and 13° C. isotherms are fairly well controlled.

## Summary

The close relationship of the early Tertiary marine faunas of the Pacific Coast, together with the subtropical character of many of them, indicates that Kamchatka could not have been within the Arctic Circle as Wegener's "floating continent" hypothesis would have it. The distribution of tropical faunas during the Eocene along the Pacific Coast of the Americas from Washington to Peru, together with similar occurrences in Java, indicates that the Eocene tropics must have included the present day tropics in the Pacific, a conclusion not in accord with Wegener's hypothesis. Further, it is noted that the marine early Tertiary faunas of the Atlantic Coast of North America do not have as close a relationship to faunas of similar age in Western Europe as would be expected if the continents were in the position postulated by Wegener, the faunal relationships not being much closer than some noted in the Recent faunas of the Eastern and Central Pacific areas. It appears that during the Cenozoic the continents and poles must have been in approximately the same positions as at present.

Based on two premises, first that, within the limits of temperatures ordinarily prevailing in marine waters, stenothermal organisms are in general much more critically limited by minimum temperatures than by maximum temperatures, and, second, that the past requirements of associations of organisms are similar to those of similar Recent associations, study of the Pacific Coast marine faunas indicates that during the Paleocene the 20° C. marine isotherm was probably north of 49° latitude and in the lower Eocene certainly was, and that since that time it has been shifted southwards, possibly with minor oscillations, until it approached its present position in the Pliocene. During the Pleistocene it oscillated both northwards and southwards, but the methods of

correlation available in this interval do not permit of a synthesis of the data into a coherent picture.

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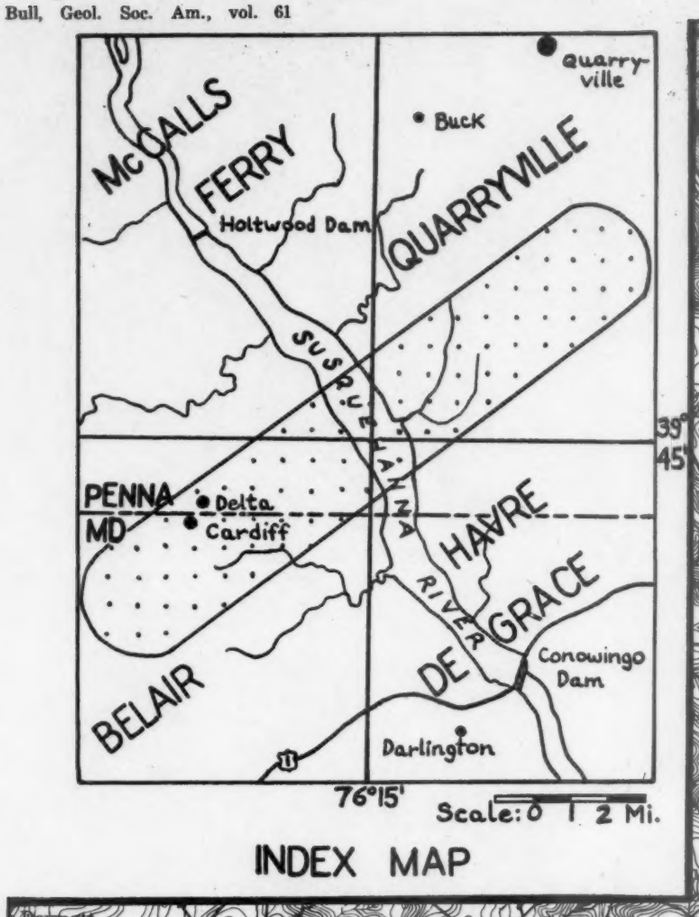



INDEX MAP

Scale 1/62500

Contour interval 20 feet.
*Datum is mean sea level.*

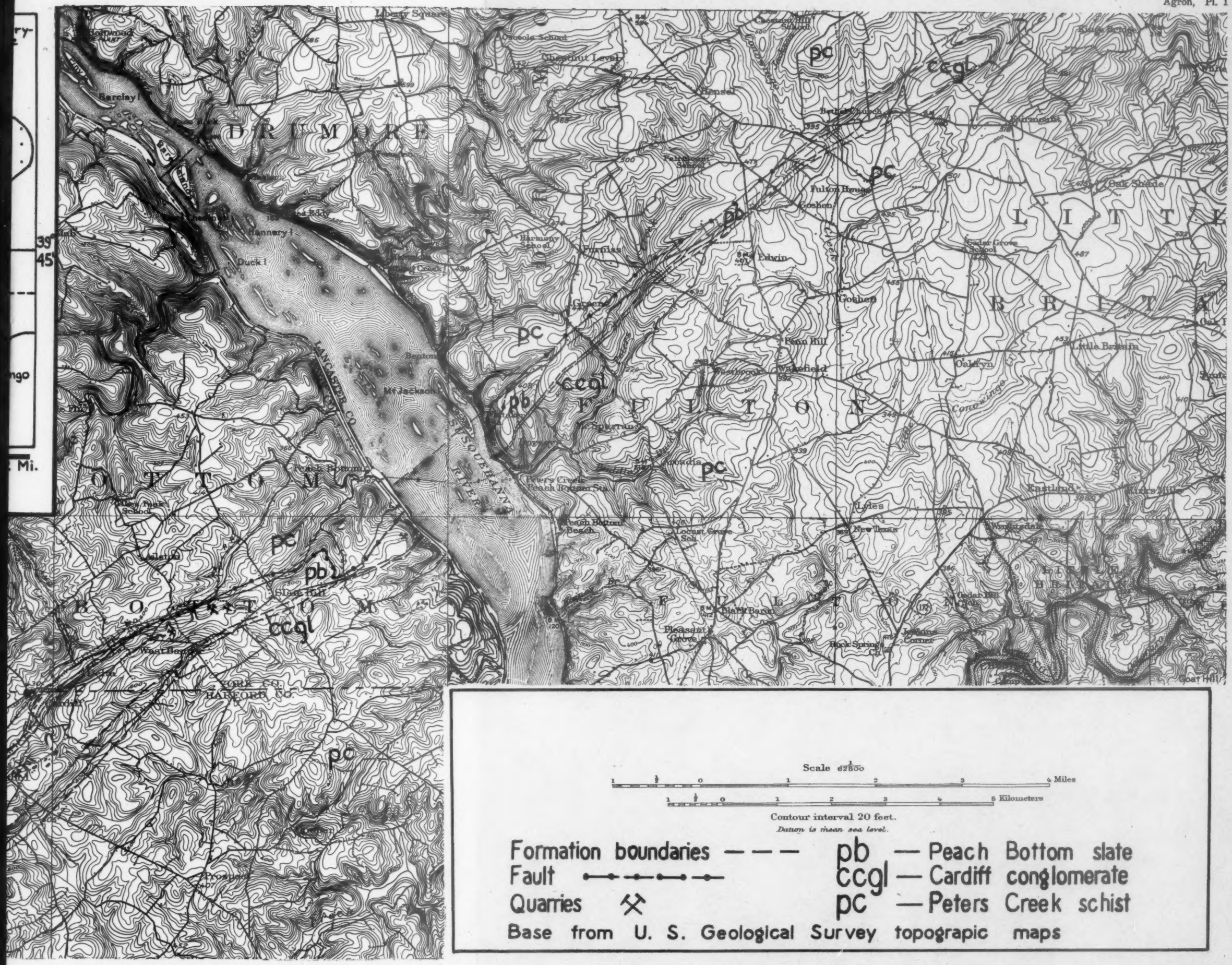
Formation boundaries — — —
Fault —•—•—•—•—
Quarries ⚒

pb —
ccgl —
pc —

Base from U. S. Geological Survey top

**MAP OF VICINITY OF PEACH BOTTOM SYNCLINE**

Index map shows location of area studied (stippled) and quadrangles.




Scale 1/62500

1 0 1 2 3 4 Miles

1 0 1 2 3 4 5 Kilometers

Contour interval 20 feet.

*Datum is mean sea level.*

Formation boundaries — — —
Fault —•—•—•—
Quarries ⚒

pb — Peach Bottom slate
ccgl — Cardiff conglomerate
pc — Peters Creek schist

Base from U. S. Geological Survey topograpic maps

## MAP OF VICINITY OF PEACH BOTTOM SYNCLINE

Index map shows location of area studied (stippled) and quadrangles.

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# STRUCTURE AND PETROLOGY OF THE PEACH BOTTOM SLATE, PENNSYLVANIA AND MARYLAND, AND ITS ENVIRONMENT

By Sam L. Agron

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## Abstract

The Peach Bottom slate lies in the center of the Peach Bottom syncline of southeastern Pennsylvania and Maryland. The slate is underlain by the Cardiff conglomerate and the Peters Creek schist. Chlorite is almost as abundant as muscovite in the slate. Chloritoid is common in the Cardiff conglomerate and in many parts of the slate belt, especially in local zones where shearing was more intense. Bedding is almost entirely destroyed, and bedding-cleavage relations cannot be used to work out the structure of the slate. The rocks show well developed flow cleavage (schistosity), fracture cleavage, joints, and lineations. Flow cleavage strikes northeast and parallels the long axis of the Peach Bottom syncline. Fracture cleavage is parallel to axial planes of drag folds and crenulations in the slate. Lineations are parallel to fold axes. They plunge generally northest to southeast at a moderate to steep angle. Steep cross joints, trending northwest to northeast, are the most abundant type. They exercise a marked control over surface drainage. Arching of fold axes, resulting in cross folds, occurred in several transverse zones at the same time that the syncline was compressed and overturned to the northwest. B-tectonites and B $\wedge$ B′-tectonites are represented. B $\wedge$ B′ is about 72°. B′ is approximately the *a* direction. Where two sets of lineations occur, there has been no uniform overprinting. Lineations are largely contemporaneous. One period of deformation, consisting of several stages, is indicated. Conclusions based on field data are borne out by petrofabric analysis.

## Introduction

### *Acknowledgments*

The writer is indebted to Ernst Cloos, who suggested the problem and offered valuable instruction and criticism. Thanks are also due to J. L. Anderson for help with petrologic aspects of the work. H. E. Vokes read the part dealing with fossils and stratigraphic relations. T. F. Bates, of the Pennsylvania State College, furnished two X-ray analyses of Peach Bottom slate. Erling Dorf, of Princeton University, James Schopf of the U. S. Geological Survey, and others have also been helpful.

### *Location of the Area*

The Peach Bottom syncline lies in the Appalachian Piedmont of southeastern Pennsylvania and Maryland. Field work was done in a

rectangular area 18 miles long, covering approximately 60 square miles, and including parts of four 15-minute quadrangles; McCalls Ferry, Quarryville, Belair, and Havre de Grace (Pl. 1).

### *Previous Work*

An early description of the slate belt and slate production in the area (Rogers 1858, p. 188–189) states that there were at that time 11 slate quarries on the Lancaster County side of the Susquehanna River and 18 on the York County side. A report of the Second Geological Survey of Pennsylvania (Frazer, 1880) supplies many interesting details of slate quarrying in each of the quarries then in operation. E. B. Mathews (1898) gives a somewhat lengthy account of the history of the slate quarrying industry in the area. Dale (1914) reviewed the Peach Bottom slate district and discussed the characteristics and mineralogy of the slate. The report of Knopf and Jones (1929) is an outstanding contribution to the geology of York and Lancaster Counties. Behre's report (1933) is the most complete structural and petrographic study of the slates. It also contains a good index map of the quarries and brief descriptions of them. The district is also discussed by Stose and Jonas (1939) and in several other brief papers.

### *Present Investigation*

The present investigation made use of the methods of modern structural geology and petrofabric techniques, not available to earlier workers. An attempt has been made to learn how slate is deformed, what deformational features and structures are developed, and which directions and orientations the structures take. A petrological study of the rocks and their metamorphism was also undertaken. Field work was done in the summers of 1947 and 1948. More than 3,000 compass readings were taken of schistosity, joints, fracture cleavage, lineations, fold axes, and bedding. The railroad cut on the east shore of the susquehanna River (the best exposure in the area) was mapped on a scale of 12½ feet to the inch. Detailed sections were also made along road cuts and other favorable exposures. Maps of several quarries were prepared. Forty-five thin sections were studied and petrofabric diagrams prepared from 13 of these.

## Rocks of the Area

### *General Statement*

The three formations recognized in the vicinity of the Peach Bottom syncline are: the Peters Creek schist, the Cardiff conglomerate, and the Peach Bottom slate. The slate and the conglomerate are more uniform lithologically than the Peters Creek schist, which includes chlorite and muscovite schists, quartzitic beds, and narrow belts of "serpentinized" rock. Several Triassic diabase dikes traverse the area, striking somewhat north of the regional schistosity.

### *Peters Creek Schist*

The Peters Creek schist is a series of interbedded chlorite-sericite and quartz-albite schists, which underlie and enclose the Peach Bottom syncline. The Peters Creek lies above the Wissahickon schist in a belt which is 6 miles wide in this area. The schist is fine-grained, greenish to grey or buff, and lithologically variable because of differences in degree of metamorphism and original mineral composition. Several quartzitic beds south of the Peach Bottom syncline compare in thickness to the Cardiff conglomerate. Some beds contain over 25 per cent feldspar, as clastic grains and as porphyroblasts. The feldspar is mostly albite, frequently about $Ab_{90}$, but oligoclase ($Ab_{80}$) and microcline are also present.

The quartz grains show various degrees of crushing and stretching (Fig. 1; Pl. 2). They frequently tail out into a mixture of very fine crush quartz, sericite, and chlorite. Some of the sandy layers carry small clastic grains of blue quartz. These show no inclusions and appear similar to other quartz grains of comparable size. Muscovite and chlorite constitute the bulk of most beds, but in some quartzitic beds they are present in amounts of less than 1 per cent. Chloritic layers are dull and dark green; muscovite-rich members are a lustrous gray or grayish brown.

Most of the outcrops in the Peters Creek

schist are composed of quartzitic layers, which frequently form knolls. Fracture cleavage, lineations, and thrusts are well developed only in the micaceous beds.

Biotite commonly occurs in very small amounts and is nearly always associated with chlorite. Alteration of biotite to chlorite is apparent in thin sections. Common accessory minerals are: magnetite, tourmaline, epidote, clinozoisite, and calcite. Pyrite, titanite, zircon, topaz, and apatite are also present, but in very small amounts.

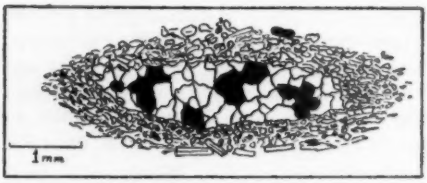


FIGURE 1. MYLONITIZED QUARTZ GRAIN IN CARDIFF CONGLOMERATE

Quartz is recrystallized and free of included matter. Matrix consists of fine crush quartz, muscovite, chlorite, pyrite, and calcite. Specimen 111, X-nicols.

The micaceous layers of the schist are finely foliated. Towards the center of the syncline, the schist is thrown into small isoclinal folds, slightly overturned to the north. Thrusts along schistosity planes are not uncommon. A short distance from the center of the syncline folding rapidly becomes less intense, and half a mile from the slate belt folds are gentle and open.

Many small quartz and calcite veins intrude the schist along foliation planes or cut across the planes at various angles (Fig. 5). Thirty feet north of the railroad bridge over Peters Creek, thin quartz veinlets are so closely spaced as to give the rock a gneissic appearance. Most of the veins are syntectonic or late syntectonic. Some are sheared out by thrusting. The effects of contact action are shown along the borders of some of the veins. In places, veins and pegmatites have caused slight addition of material to the schist.

The Peters Creek schists contains several narrow "serpentinized" belts which parallel the regional schistosity. One such belt, which is not over 500 feet wide, extends 2 miles southwesterly from the state line at Cardiff, Maryland. Another belt, 5 miles long and 300 feet wide, extends southwest from a point about a mile northwest of Pylesville. Several other "serpentinized" bodies in the Peters Creek schist are indicated by Knopf and Jonas (1929) and Stose and Jones (1939). X-ray analysis of one specimen indicated that it contained predominantly chrysotile plus a chlorite (T. E. Beukelman, written communication).

### *Cardiff Conglomerate*

The Cardiff conglomerate overlies the Peters Creek schist and encloses the Peach Bottom slate. It extends northeast from a point half a mile east of Pylesville, Maryland, to about three-fourths of a mile southwest of the West Branch of Octoraro Creek. The formation is, in places, a conglomeratic quartz schist containing sheared, elliptical quartz pebbles up to 4 inches long. The matrix consists of fine-grained quartz, plagioclase, muscovite, chlorite, and chloritoid. Accessory minerals are: magnetite, pyrite, calcite, tourmaline, epidote, zircon, and titanite. Carbonaceous matter is locally present. Where shearing has been very intense, as near the northeast end of the syncline, a mile northwest of Fairmont, there are films of highly lustrous graphite. Some thin sections show brown remnants of biotite where larger flakes have altered to chlorite. Chlorite and muscovite flakes are often wrapped around lenticular quartz pebbles. They are also intimately intergown with crush quartz at the tapered ends of pebbles. The crush quartz particles average about 0.04 mm. in size.

The contact between the Cardiff conglomerate and the Peters Creek schist is rarely exposed. Along the east shore of the Susquehanna River there is a weathered zone 6 inches wide, at the contact. Here, as well as on the west side of the river, the contact is conformable.

The slate-conglomerate contact shows several feet of thin alternating slaty and quartzitic beds. The conglomerate is 68 feet thick in the section along the railroad out on the east bank of the Susquehanna River. Stose and Jonas (1939) gives a thickness of 72 feet for a section three-quarters of a mile southeast of Whiteford, Maryland. These values do not represent stratigraphic thicknesses because the formation has been thinned by shearing.

Schistosity is parallel to bedding and forms straight smooth surfaces. The conglomerate is well-jointed. Rhomboid blocks are often formed by bedding planes intersecting with two sets of transverse joints inclined 60° to each other. Fresh exposures of the conglomerate show lenticular white quartz pebbles when viewed normal to the schistosity. They appear as rounded blotches of quartz when viewed in the schistosity plane. The groundmass is light gray and fine-grained. Muscovite and chlorite give a lustrous appearance to schistosity planes. A faint lineation parallel to *b* is visible where micaceous minerals are sufficiently abundant. Weathered outcrops are buff-colored, reddish, and occasionally greenish, presumably where the chlorite content is high. On weathering, the mortised and sheared quartz pebbles crumble to a fine sand.

STRETCHED PEBBLES. Several measurements of the dimensions of stretched pebbles in the conglomerate were made, but a systematic study of pebble stretching was not undertaken. Axes *a* and *b* of the pebbles lie in the schistosity plane; *b* is parallel to the regional fold axes and is the axis of maximum elongation. The *c* axis is perpendicular to *a* and *b*.

A quartz pebble from the abandoned quarry in the Cardiff conglomerate three-fourths of a mile southeast of Whiteford, Maryland is 1.7 x 3.7 x 0.9 inches in size. The ratio *a*:*b*:*c* is 2:4:1. In an outcrop on the east side of the dirt road three-fourths of a mile north of Peach Bottom station in Lancaster County, a strongly sheared pebble has axial ratios of 6:11:1. Over the entire area, the range of a:b:c was from 2:2.4:1 to 10:18:1. Cloos (1947, p. 852) discusses stretched pebbles and cites several examples from the literature. The range given was similar to that found in this area; *b* to *c* ratios of 1:1 to 2:1 were reported in quartz, and ratios of 18:1 in schistose greywacke and shales. The shape of the deformed pebbles depends on the intensity of deformation and on the material. As the quartz pebbles in the Cardiff conglomerate are all of the same material, the great range in stretching indicates that deformation throughout the Peach Bottom syncline was not of uniform intensity.

### *Peach Bottom Slate*

*General statement.*—The Peach Bottom slate extends along the strike of the Peach Bottom syncline for 13½ miles. In York County, Pennsylvania, and Harford County, Maryland, it forms "slate ridge", which is half a mile wide and about 150 feet above the surrounding country. In Lancaster County, the slate belt thins greatly and exercises very little control on the topography. The Cardiff conglomerate underlies the slate and surrounds it everywhere except on the northwest where, for a distance of 9 miles, from a point just northeast of Delta to one southeast of Fairmount school in Lancaster County, the slate is faulted against the Peters Creek schist.

*Slate-conglomerate contact.*—The contact between the slate and the conglomerate shows several feet of alternating slate and quartzitic beds. Behre (1933, p. 364) observed thin quartzitic interbeds, rarely more than half an inch thick, in several old quarries half a mile south of Delta, and also on the north limb of the syncline along the east bank of the Susquehanna River.

The contact on the south side of the syncline, along the Susquehanna River, is sharp. However, the incompetent slate is strongly contorted and lineated adjacent to the contact. It is also intruded by thick quartz veins. Along the north-south dirt road three-eights of a mile east of the river, the contact between the two formations consists of interbedded slate and quartzitic beds. A 6-inch bed of quartzitic slate and two 6-inch beds of schistose slate are separated by thinner beds of quartzitic material. One and a fourth miles to the northeast, in the outcrop along the road a half mile southeast of Greene, the contact shows 1½ feet of interbedded quartzitic layers and slate. Two zones of finely intercalated slate and quartz are present in the northernmost 50 feet of the slate exposure along the railroad cut on the east bank of the Susquehanna River. The larger of these zones is about 6 feet wide. The quartzitic layers are as much as half an inch thick. The slaty layers, along which most of the shearing took place, range in thickness from mere films to a quarter of an inch.

*Quartzitic lenses in slate.*—A gray quartzitic lens, about 25 feet thick crops out in a bench 80 feet south of the southeast rim of Kell quarry, which is about a quarter of a mile southeast of the Funkhouser plant. This lens appears to be

near the south contact of the slate. Another lenticular gray quartzitic body, up to 2 feet thick, was exposed in overburden stripping operations on the ledge southwest of the Funkhouser quarry. Numerous thin sandy intercalations were found in the slate a few yards north

TABLE 1. X-RAY ANALYSES OF SLATE

| Mineral | Specimen No. 13 | Specimen No. 35 |
|---|---|---|
| Quartz | 25 | 0 |
| Feldspar | 4 | 4 |
| Pyrite | 5 | 0 |
| Sericite-illite | 35 | 40 |
| Chlorite | 24 | 42 |
| Calcite | 5 | 8 |
| Dolomite | 3 | 6 |
| Total | 101 | 100 |

of this lense. The contact between the quartzitic layers and the slate is interpreted as bedding. It is parallel to the slaty cleavage.

The quantity of commercial grade slate in the district is limited by numerous zones of intense deformation in which the slate is strongly folded, sheared, and crushed. These crushed zones contain many quartz veins, called "flint rock" by the quarrymen. At the southwest end of the syncline, south of Whiteford, Maryland, joint planes are spaced too closely together to permit recovery of large squares of slate suitable for blackboards.

*Composition of the slate.*—Quartz, muscovite, chlorite, and carbonaceous matter (locally graphitic) are the chief mineral constituents of the slate. Chloritoid is common. Andalusite is abundantly present, according to Dale (1914 p. 112). Minor constituents and accessory minerals are: pyrite, magnetite, tourmaline, rutile, titanite, zircon, and zeolites. Albite is also present, but very difficult to distinguish from minute quartz grains. X-ray analyses indicate the presence of calcite and dolomite, but they cannot be seen in thin sections. Many of the minerals of the groundmass are about 0.004 mm. in size, and the largest grains rarely exceed 0.2 mm. Accurate modal analyses, therefore, cannot be made with the petrographic microscope. Much better quantitative results can be obtained by X-ray analysis.

The slates of northeastern Pennsylvania were recently studied at the School of Mineral Industries of the Pennsylvania State College. In addition to standard petrographic studies, the slates were subjected to chemical and thermal analyses, and X-ray and electron microscope examination. The analyses indicate that a mineral of the illite group forms the bulk of the micaceous minerals in the Lehigh-Northampton districts (Bates, 1947). Mineral orientation in the more metamorphosed "hard slate" is much better than in the less metamorphosed "soft slate".

*X-ray analyses of Peach Bottom slate.*—Dr. T. F. Bates of the School of Mineral Industries, The Pennsylvania State College, very kindly made X-ray diffraction patterns of two specimens of Peach Bottom slate submitted by the writer. An "educated guess" of the percentages of the minerals present in these samples is presented in Table 1.

Comparison of these analyses with 20 analyses of slates from Pennsylvania, Virginia, and Vermont (The Pennsylvania State College. Mineral Industries Experiment Station, 1947, Fig. 14) shows that the two samples of Peach Bottom slate are considerably lower in quartz and pyrite, lower in feldspar and calcite, considerably higher in sericite-illite, and very much higher in chlorite. The extraordinarily high chlorite content is perhaps the outstanding characteristic of Peach Bottom slate.

Behre (1933, p. 375) points out that the Peach Bottom slate has a higher silica content than other Pennsylvania slates. As its quartz content is low, much of the silica must have entered recrystallized silicates such as chlorite and sericite-illite. The two samples of Peach Bottom slate, especially No. 35, do show an unusually high percentage of micaceous minerals. Chemical analyses also indicate that the Peach Bottom slate contains more alumina and iron, and less lime, magnesia, carbonate, pyrite, and water than other Pennsylvania slates.

## MINERALOGY

### *General Statement*

Slightly more than 20 species are represented. The most abundant minerals are quartz, feldspar, muscovite, and chlorite. Common accessories are chloritoid, pyrite, epidote, magnetite,

calcite, tourmaline, and, sometimes, biotite. Titanite, rutile, zircon, andalusite, topaz, and garnet are not common. Chrysotile is locally abundant in belts in the Peters Creek Schist. Zeolites occur in the slate, but they are rare.

### *Quartz*

Almost all the quartz is recrystallized. Crush quartz is fairly abundant. Needle quartz is sometimes present in veins, but it is rare. Peripheral growth quartz is the first indication of deformation, and forms even before the grains show strain shadows (Fellows, 1943). The fact that this type of quartz has not been observed by the present writer indicates that nowhere was the intensity of deformation extremely slight. Outlines of grains and pebbles can frequently be seen, but the grains are stretched, crushed, and recrystallized. Some pebbles show a peripheral strained area about an unstrained center. In other grains recrystallization has begun around the edges of a strained central portion.

Quartz grains and pebbles were crushed and sheared in the plane of the schistosity. Subsequent recrystallization of much of the crush quartz produced a granoblastic texture. Some recrystallized grains show straight polygonal boundaries. Other clusters of crush quartz are only partially recrystallized. Intermediate stages are indicated by the merging of individual crush quartz grains and by sutured boundares within the cluster. Some of the recrystallized quartz has subsequently developed strain shadows.

Figure 2-a shows the direction and the degree of plunge of *c*-axes in different parts of a strained quartz grain and also four axes in adjacent grains. The strained quartz grain shows undulose extinction. Figure 2-b shows the *c*-axes in a quartz grain which has begun to recrystallize. The *c*-axes in the strained portion of the grain plunge 21°–32° towards the lower right, but show no uniformity of direction in the recrystallized parts of the grain. Hietanen (1938, p. 33–34) attributes the slight differences in the orientation of *c*-axes in quartz showing undulose extinction to folding or crumpling of the basal pinacoid.

Quartz in syntectonic veins is completely recrystallized and shows a granoblastic texture. In late syntectonic veins the quartz has strong strain shadows (Pl. 2, figs. 4, and 5), and some grains are fractured parallel to the strain shadows. A few veins contain needle quartz. Lamellae are present in some vein quartz, but they are not common.

Small postkinematic veinlets carry clear unstrained quartz. Such late quartz replaces calcite in the Cardiff conglomerate. Quartz crystals in veinlet are about 0.2 mm. in size. In the slate the quartz grains are much smaller, averaging about 0.02 mm.

Detrital grains of blue quartz, about 2 mm. in diameter, are common in many beds in the Peters Creek schist. They are fairly widely distributed in the rocks of the Appalachian Piedmont.

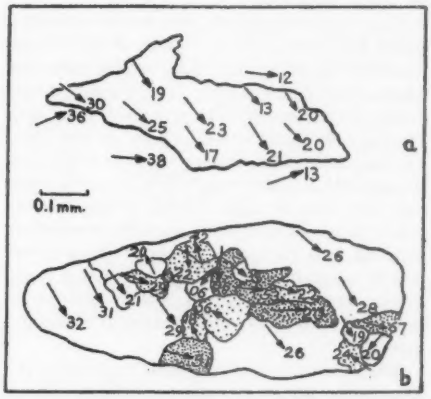


FIGURE 2. ORIENTATION OF *c*-AXES (a) IN STRAINED QUARTZ GRAIN, (b) IN RECRYSTALLIZED QUARTZ GRAIN

### *Feldspar*

Albite is the common feldspar. In some arkosic beds it constitutes about 28 per cent of the rock, occurring as porphyroblasts up to 1.5 mm. across and as rounded detrital grains of smaller size. Very small albite grains in the groundmass of the schist average about 0.05 mm. in size. Some grains show fine albite twinning, while others show complex twinning.

Extinction angles of albite twins, measured in the [010] zone, usually indicate a soda-rich albite, frequently $Ab_{98}$. The extinction angles also indicate the presence of oligoclase in lesser amounts. Measurements of 2 V were made on the universal stage. All methods showed a preponderance of albite, and some oligoclase and potash feldspar.

Inclusions in albite porphyroblasts consist of quartz and muscovite, and a little epidote. In some grains the inclusions are oriented parallel to crystallographic planes, in other grains no orientation is apparent. Ingerson (1936) considered that mica inclusions which were parallel to crystallographic planes of albite in the Port Deposit granite were formed later than the albite, perhaps by hydrothermal agency.

Hietanen observed albites in the Wissahickon schist which showed clear margins around central areas containing epidote inclusions (Cloos and Hietanen, 1941, p. 104). A few such albites were

also observed by the writer in the Peters Creek schist (Pl. 2, fig. 6). Hietanen postulated that the material which formed the epidote inclusions was trapped in the albite grains during the early stage of formation. At a later stage, when the clear rim was formed, the material which formed epidote was no longer available.

Some of the clear albite rims around sericitized grains have grown at the expense of surrounding quartz grains. A few large flakes of late muscovite project into the clear albite rims, suggesting contemporaneous growth. The sericitized cores may represent weathered detrital grains, or they may, perhaps, indicate a decrease in temperature and metamorphic activity, which later increased when the clear rims of albite formed. At least two increases in temperature in the Wissahickon schist are established on the basis of zoning in albites (Cloos and Hietanen, 1941, p. 106). The increases in temperature were followed by periods of low-grade metamorphism, as shown by chloritization of biotite and garnet.

The intensity of metamorphism varied considerably in the limited area about the Peach Bottom syncline. Although most of the feldspar in the formations is authigenic, there is an abundance of detrital plagioclase where metamorphism was locally less intense. A few detrital grains are fractured, the several parts indicating rotation.

Microcline is porphyroblastic and clearly younger than adjacent quartz. Orthoclase is very scarce, as most of the available potash went into the formation of muscovite.

### *Muscovite*

Muscovite is abundant in every formation. It occurs in fibrous masses and well-formed laths lying in the schistosity, as well as large porphyroblastic flakes growing across the schistosity. The smaller laths are the sericitic variety. Their size in the slates is about 0.025 mm.

Muscovite forms very early in metamorphism. Small scales of sericite make their appearance in some sandy layers in slate which has been so little metamorphosed that much of the carbonaceous and argillaceous matter is not recrystallized. As metamorphism progressed, fibrous muscovite became less abundant and well-formed laths more prominent.

Fine intergrowths of chlorite and muscovite are common. They are often intimately associated with crush quartz. Some flakes of muscovite grow on faces of pyrite cubes.

Sericite laths frequently show a faint yellow to green pleochroism. Optical data, as determined from speciment 170, indicated 2V ca. 33°; $\beta = 1.605$. This indicates a high percentage of the phengite molecule. Thus, this specimen is slightly higher in iron and considerably higher in magnesium than the muscovite from the Wissahickon schist at Safe Harbor, which had a $2V = 40° - 44°$ and $\gamma = \beta = 1.602$ (Cloos and Hietanen, 1941, p. 98). The muscovite from specimen 170 appears to be similar to that found by Barth in Dutchess County, New York, in which $2V = 30° - 40°$ and $\gamma = \beta = 1.600+$ (Barth, 1936).

Specimen 170 was taken half a mile southeast of the slate belt and may not be representative for the area as a whole.

### *Chlorite*

There is much chlorite in the Peters Creek schist and the Peach Bottom slate, but only small amounts appear in the Cardiff conglomerate. It occurs in flakes and laths up to 0.2 mm. long, as well as in fibrous aggregates intergrown with muscovite. Quartz and calcite veins usually contain small amounts of chlorite.

The chlorite is pleochroic, green to yellow or colorless. In many places it appears to be altering from biotite, sometimes with abundant associated magnetite. Chlorite flakes in the Cardiff conglomerate may show pleochroic halos surrounded by brownish biotitic matter. Some chlorite shows on magnetite crystals, the two minerals probably being contemporaneous. In calcite veins, the chlorite appears to be older than calcite. Most of the chlorite formed during the recrystallization of the rocks, but some is secondarily derived from the alteration of biotite.

Optical data were determined for six specimens. $n_g$ was measured on a flat stage, with the use of oils. The value of $n_\gamma - n_\alpha$ was determined with the Berek compensator; with the thin sections oriented on the universal stage in the position yielding maximum birefringence.

The characteristics of six chlorite specimens are listed in Table 2. Well No. 1 was a churn well in the Peters Creek schist on the north side of the road, about 500 feet east of the Funkhouser plant near Delta. Well No. 5 was 50 feet south of the road, opposite the office of the Funkhouser Company. These wells are immediately north of the fault contact between the slate and the schist.

The varieties of cholrite represented are thuringite, ripidolite, and aphrosiderite. Clinochlore, reported to be the variety of chlorite in the Peters Creek schist by Knopf and Jonas (1929, p. 36), was not observed, and is perhaps not to be expected in view of the high Fe:Mg ratios. These ratios range from 5:5 to 9:1. The Al:Si ratios range from 6:4 to 8:2. Thus the chlorite is consistently high in iron and alumina.

### *Chloritoid*

Chloritoid, fairly abundant in the slate and somewhat less abundant in the conglomerate, occurs in laths or tabular crystals ranging in length from about 0.04 mm. to 0.4 mm. The length is approximately six times the width. The laths may appear as scattered porphyroblasts (Pl. 3, fig. 1), they may form rosettes or sheaf-like aggregates, or they may appear as a mesh of crystals with minute interstitial grains of quartz and muscovite. This arrangement resembles the diabasic texture in igneous rocks.

Although chloritoid crystalloblasts transect flow cleavage at all angles, many laths lie parallel to flow cleavage. Some, together with muscovite flakes, are wrapped around quartz pebbles. Chloritoid laths commonly show cross partings and ragged ends. Many show inclusions of quartz and carbonaceous matter. Some colorless "skeleton" crystals with sharp dark outlines seem to be just forming from the argillaceous matrix.

Most grains are moderately pleochroic with $\alpha$ = gray-green or brownish-green, $\beta$ = slate blue, $\gamma$ = yellow to colorless. $\alpha = 1.725$; $\beta = 1.731$; $+ 2V$ = ca. 45°; $Z \wedge c = 18°$. Some chloritoid shows an extinction angle of 23°, suggesting ottrelite. Polysynthetic twinning is not uncommon. There are usually two individuals, rarely as many as four.

### *Biotite*

Biotite is present in very small amounts both north and south of the slate belt. It is also found in some calcite-quartz veins in the schist. Biotite occurs in shreds, frequently streaked with chlorite, and in laths about 0.04 mm. long. There are also large ragged porphyroblasts of later biotite cutting across the schistosity. The mineral almost always shows alteration to chlorite. Some large areas of green chlorite still contain brown patches of high relief, showing typical biotite pleochroism. Both brown and green biotite are present. Some calcite veins contain the green variety, which is pleochroic green to yellow.

### *Epidote minerals*

Pistacite and clinozoisite are common accessories in the Peters Creek schist. They are somewhat less common in the Cardiff conglomerate. The epidote minerals occur in rounded grains and elongated crystals up to 0.3 mm. x 0.2 mm. The core in many of the grains is slightly pleochroic, yellow to colorless, and is surrounded by a non-pleochroic rim. This suggests that the iron in the vicinity of the crystal went into the formation of yellow pleochroic pistacite. The zone surrounding the crystal was leached of iron and, as growth continued, a colorless rim of non-pleochroic, iron-poor clinozoisite formed.

A similar instance of metamorphic differentiation according to the concretion principle is cited by Turner (1948a, p. 138) in the low-grade quartz-albite-epidote-muscovite schists of Otago, New Zealand. These schists were found to contain 0.1 per cent MnO, all of which is accounted for by small garnet crystals carrying 20 to 30 per cent MnO.

Some of the epidote grains lie parallel to the schistosity. A few grains are badly fractured, indicating that they are not postkinematic.

The schist adjacent to some calcite veins shows strong epidotization. The contact zone consists of a mass of fine-grained quartz epidote, sericite, chlorite, and tourmaline.

### *Pyrite*

All formations contain small amounts of pyrite as idiomorphic porphyroblasts or as irregular disseminations. It is also common in late quartz veins in the slate. A postkinematic age is indicated for these quartz-pyrite veins by the fact that comb-like quartz crystals growing normal to their walls have not been disturbed.

Some zones in the slate show lenticular eyes of

TABLE 2. VARIETIES OF CHLORITE

| Specimen number | Formation | Sign of 2V | $\beta$ | $\gamma-\alpha$ | Name (Winchell 1932, p. 279) | Fe:Mg | Al:Si |
|---|---|---|---|---|---|---|---|
| 111 | Ccgl. | − | 1.666 | 0.006 | Thuringite | 9:1 | 8:2 |
| 160 | Pc | + | 1.624 | 0.004 | Ripidolite | 5:5 | 8:2 |
| 170 | Pc | − | 1.632 | 0.002 | Aphrosiderite | 6:4 | 6:4 |
| 171 | Pc | − | 1.644 | 0.002 | Aphrosiderite | 7:3 | 8:2 |
| Well 1 | Pc 45 ft. | + | 1.614 | | | | |
| Well 5 | Pc 220 ft. | + | 1.632 | | | | |

No. 1 well and No. 5 well were churn wells drilled in the Peters Creek schist in an unsuccessful search for water. The beta index and the sign were determined for the chlorite from these well cuttings. 45 ft. and 220 ft. are the depths at which the samples were taken.

quartz which are divided by transverse partitions of pyrite. The pyrite septa parallel fracture cleavage and are about 0.2 mm. thick. The lenticular bodies of quartz are elongated in the direction of slaty cleavage. Individual quartz crystals within the eyes grow parallel to the length of the eye and normal to the transverse pyrite partition. Pabst (1931) described somewhat similar feather quartz in the "pressure-shadows" of pyrite porphyroblasts in the Calaveras slates. Plate 3, figure 2, shows late porphyroblastic pyrite replacing quartz along finger-like projections which grow normal to the cube faces. Under reflected light, this porphyroblast shows two growth stages. The later stage is rotated 7° to the north.

One porphyroblast, 0.3 mm. in diameter, is a rosette of minute cubes (Fig. 3). Cubic porphyroblasts of pyrite, as large as half an inch on a face, are abundant in the schist along the east bank of the Susquehanna River, about 1300 feet northwest of the railroad bridge over Peters Creek. The cubes do not present any faces parallel to the schistosity. Some thin sections show red translucent pseudomorphs of iron oxide after pyrite.

### *Magnetite*

Magnetite is a common accessory mineral. Locally it comprises more than 2 per cent of the schist. It occurs in euhedral to anhedral grains up to 0.75 mm. across, and commonly, in granular aggregates. Some large crystals appear to have been crushed and spread out along schistosity planes. Chlorite sometimes grows on magnetite crystals, the two minerals are then probably contemporaneous.

### *Calcite*

Calcite is present in veinlets and as irregular grains along schistosity planes. It is usually later

PLATE 2.—PHOTOMICROGRAPHS

FIGURE 1. STRETCHED AND CRUSHED QUARTZ GRAINS IN CARDIFF CONGLOMERATE. Tailed intergranular area consists of fine strands of quartz and muscovite. Specimen 119. Plain light.

FIGURE 2. SAME AS FIGURE 1, BUT WITH CROSSED NICOLS. Quartz grains are completely recrystallized and are surrounded by crush quartz and muscovite.

FIGURE 3. INCOMPLETELY RECRYSTALLIZED QUARTZ IN CARDIFF CONGLOMERATE. Granoblastic texture is being developed. Sutured grain boundaries indicate readjustment is not complete. Specimen 104. X-nicols.

FIGURE 4. LATE SYNTECTONIC QUARTZ VEIN IN SLATE. Shows badly strained quartz (left) and sutured quartz which has begun to recrystallize (right). Muscovite strand extends from top to bottom near middle of picture. Specimen 35. X-nicols.

FIGURE 5. STRAINED QUARTZ IN LATE SYNTECTONIC QUARTZ VEIN IN PETERS CREEK SCHIST. Lamellae, oriented perpendicular to the strain shadows, can be discerned faintly. Thin strands of muscovite follow two sets of fractures which are about 50 degrees apart. Note sutured boundaries. Specimen 173. X-nicols.

FIGURE 6. ALBITE PORPHYROBLAST IN PETERS CREEK SCHIST SHOWING CLEAR ALBITE RIM AROUND SERICITIZED CORE. Muscovite lathes follow schistosity, which trends from upper left to lower right. Quartz shows granoblastic texture. Specimen 104. X-nicols.

PLATE 3.—PHOTOMICROGRAPHS

FIGURE 1. CHLORITOID PORPHYROBLASTS IN SLATE. Note rosette near upper left. Specimen 44. Plain light.

FIGURE 2. PYRITE PORPHYROBLAST REPLACING QUARTZ IN CARDIFF CONGLOMERATE ALONG FINGER-LIKE PROJECTIONS NORMAL TO THE CUBE FACES. Growth is rectilinear. The lower left portion of the porphyroblast is inclined 7 degrees to the major mass. The crystal shows two growth stages in reflected light. The later stage, which contains the long projections, is rotated 7 degrees counter-clockwise (to the north) with respect to the first stage. Quartz, chloritoid, and some calcite are also present. Specimen 111. Plain light.

FIGURE 3. NATROLITE SURROUNDED BY RIM OF QUARTZ (WHITE) IN CAVITY IN SLATE. The long dimension of the cavity is 0.35 mm. Specimen 2. Plain light.

FIGURE 4. CHLORITE FORMING ALONG CONTACT BETWEEN QUARTZ VEIN AND SLATE. Wedge of chlorite extends a short distance into the slate. Specimen 56. Plain light.

FIGURE 5. FRACTURE CLEAVAGE IN SLATE. Fracture cleavage forms in micaceous layers, but does not extend into quartzitic layers. Slaty cleavage planes are displaced due to shearing along fracture cleavage. Specimen 55. Plain light.

FIGURE 6. DETAIL OF FRACTURE CLEAVAGE IN SLATE. Dark areas are carbonaceous matter and some muscovite. Light areas are mostly quartz, and some muscovite and chlorite. Note thinning of limbs along fracture cleavage planes. Specimen 37c. Plane light.

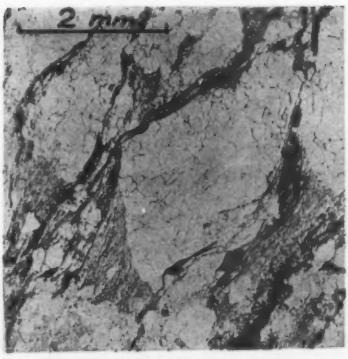


FIGURE 1

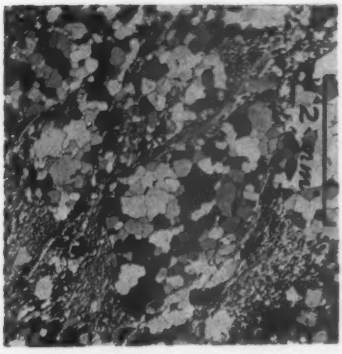


FIGURE 2

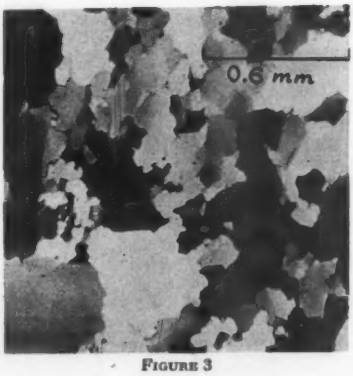


FIGURE 3

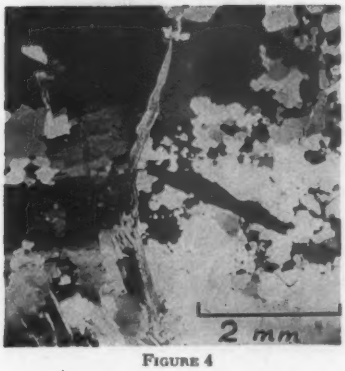


FIGURE 4

FIGURE 5

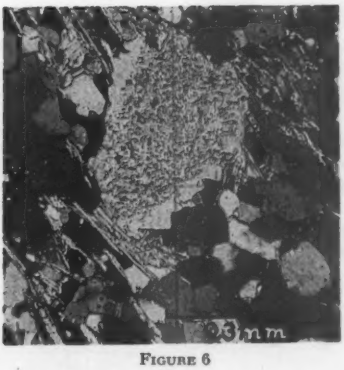


FIGURE 6

PHOTOMICROGRAPHS

FIGURE 1

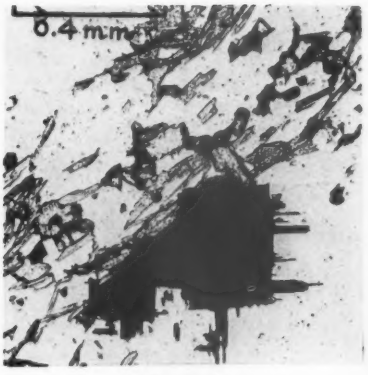


FIGURE 2

FIGURE 3

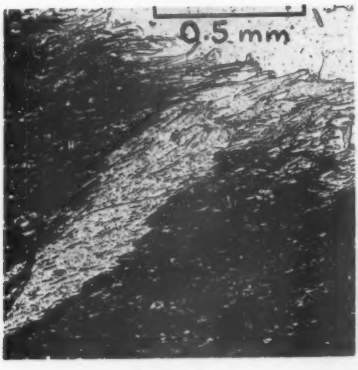


FIGURE 4

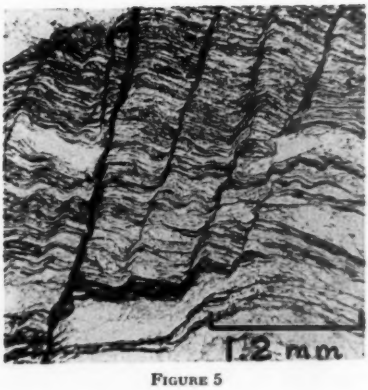


FIGURE 5

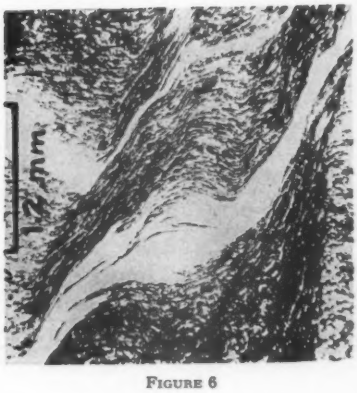


FIGURE 6

PHOTOMICROGRAPHS

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

TECTONIC STRUCTURES

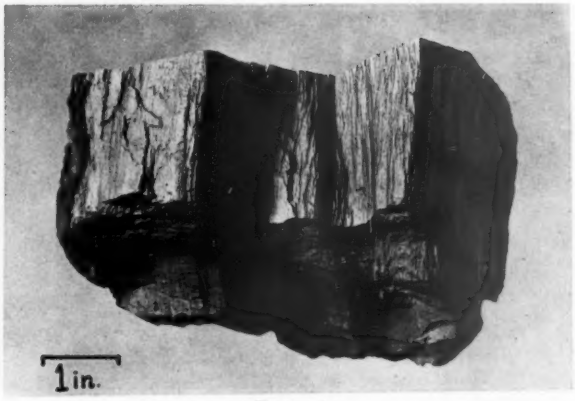


FIGURE 1

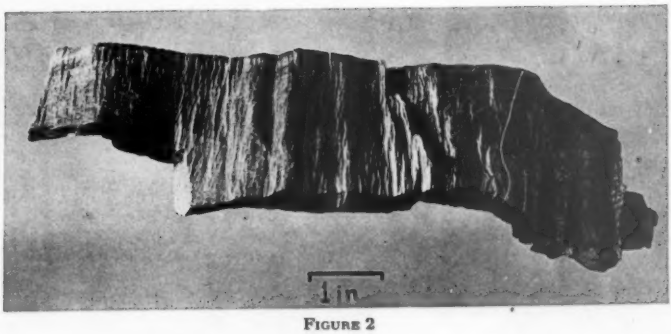


FIGURE 2

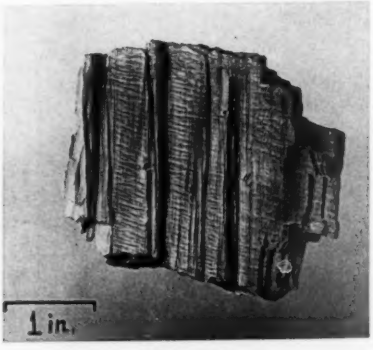


FIGURE 3

FIGURE 4

LINEATIONS IN SLATE

than quartz, which it replaces along cracks and grain boundaries leaving island-like blebs of quartz surrounded by advancing calcite. In the Cardiff conglomerate, however, very late clear quartz can be seen replacing calcite.

Some calcite grains present good cleavage boundaries to chlorite flakes, which they replace.

Calcite is usually late, but some of it formed before deformation had ceased altogether. This is shown by slightly displaced cleavage traces, which indicate rotation.

### *Tourmaline*

Tourmaline is much more abundant in the schist than in the slate. Usually minute grains or prisms may reach as much as 0.3 mm. in length. They may lie in the schistosity plane, sometimes with their *c* axis parallel to the fold axis. A few larger grains are crushed and form clusters, the cracks of which are healed by quartz. Tourmaline is present in sandy or gritty beds as well as in argillaceous layers. It has not been observed in any of the quartz or calcite veins. Most of the tourmaline is greenish-brown, but some is pinkish. There are several pleochroic formulae: E to O colorless to greenish-brown, pink to black, and colorless to yellow. The first two formulae suggest schorlite, and the third suggests dravite. Some prisms show a longitudinal zoning of colors, brown at one end, colorless in the middle, and blue-green at the other end.

### *Titanite*

All the formations contain very small amounts of accessory titanite. The grains are pink and less than 0.02 mm. in size.

### *Rutile*

Rutile occurs in the slate as irregular concretionary granules composed of tiny crystals which are rosetted at 60° (Behre, 1933). The grains are less

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PLATE 4.—TECTONIC STRUCTURES

FIGURE 1. IRREGULAR BEDDING CONTACT (dashed line) IN PETERS CREEK SCHIST. Light quartzitic schist above, dark chloritic schist below bedding plane. Cleavage dips steeply to the right (southeast). Railroad track in lower right corner gives scale.

FIGURE 2. COMPLEX THRUST FAULT IN PETERS CREEK SCHIST. Along railroad cut about 1350 feet northwest of Peters Creek. View looking northeast.

FIGURE 3. SMALL ISOCLINAL FOLD IN PETERS CREEK SCHIST (Outlined by dashed line). Pick end of hammer points toward bottom of fold axis, which plunges steeply to the east.

FIGURE 4. FOLDED SLATE IN JOHNSON QUARRY CUT BY REVERSE FAULT WHICH DIPS STEEPLY TO NORTH (PARALLEL TO HAMMER HANDLE). Fold axis plunges toward observer (east). White areas at left and upper right corner are quartz veins.

FIGURE 5. OFFSETS TO NORTHWEST ALONG FRACTURE CLEAVAGE IN SLATE ABOUT 2000 FEET NORTHWEST OF PETERS CREEK. Slaty cleavage dips steeply southeast and is transected by gently-dipping fracture cleavage in a lineation plunging gently to the left. View looking southeast.

PLATE 5.—LINEATIONS IN SLATE

FIGURE 1. FOLDED SLATE SHOWING LINEATIONS PARALLEL TO FOLD AXES. Lineations form where slaty cleavage intersects fracture cleavage (dark planes near middle and at right). Limbs of folds are sheared and crests are thickened.

FIGURE 2. CORRUGATED SLATE RIBBON FROM KELL QUARRY. Lineations are parallel to fold axes and are formed by intersection of fracture cleavgae, which is subparallel to axial plane, and slaty cleavage, which appears corrugated. Upper and lower edges of the specimen are bounded by *ac* joints.

FIGURE 3. SLATE SHOWING TWO SETS OF LINEATIONS INTERSECTING AT AN ANGLE OF 73°. The gently plunging lineations are formed by the intersection of fracture cleavage ($S_3$) with slaty cleavage ($S_2$), and they are parallel to B. The steep lineations are parallel to B′ and are formed by the intersection of a fracture cleavage ($S_4$) with slaty cleavage. The three S-planes do not all intersect in a B axis. BB′ = 73°. The horizontal linetions are slightly older than the steep ones, as they are transected by the latter. A few steep lineations, however, are early and are transected by horizontal ones. The two sets are thus largely contemporaneous and were formed by the same deformation. Several small pyrite porphyroblasts are visible.

FIGURE 4. THREE SETS OF LINEATIONS IN SLATE LEDGE ABOVE NORTHWEST RIM OF KELL QUARRY. The most prominent set of lineations plunge 75° to the south-southwest. Another set, to the left of hammer head, plunges 36° to the southwest. The third set, parallel to the hammer handle, is fainter and plunges 15° to the east-northeast.

than 0.1 mm. in diameter and are best seen in sections parallel to slaty cleavage.

### *Andalusite*

Dale (1914, p. 111) considered the abundance of minute crystals of andalusite the most conspicuous

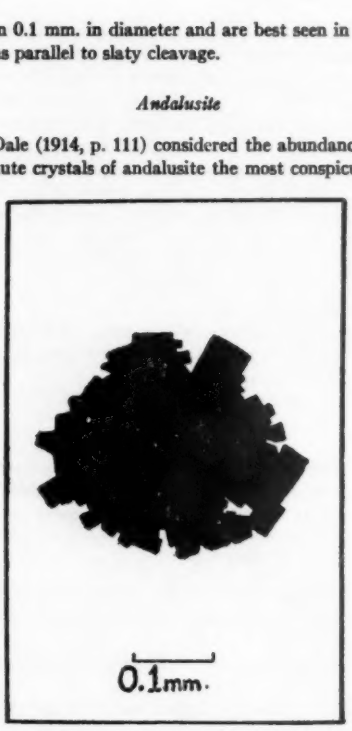



FIGURE 3. PORPHYROBLASTIC ROSETTE OF PYRITE CUBES IN PETERS CREEK SCHIST
Specimen 172.

feature of the Peach Bottom slate. In the thin sections examined by the writer, andalusite was not very abundant. Several square grains as large as 0.063 mm. were seen. They are pleochroic, pink to green, and contains dark inclusions (carbonaceous matter?). One porphyroblast, in a layer of chloritoid and carbonaceous matter, appears to have pushed aside some of the chloritoid laths.

### *Zircon*

Occasional detrital grains of zircon are found. They attain 0.25 mm., are usually rounded, but sometimes prismatic. One prismatic crystal was elongated in the schistosity plane and tapered to somewhat frayed ends (Fig. 4).

### *Natrolite*

Natrolite was found in very small cavities in a specimen of slate from Kell quarry. It forms colorless to yellowish rounded masses or spherulites which are composed of almost sub-microscopic fibers. The spherulites have a radius of 0.05 mm. Under crossed nicols they yield a dark cross which resembles that produced by uniaxial minerals in conoscopic observation. Those fibers which, at any time during rotation of the microscope stage,

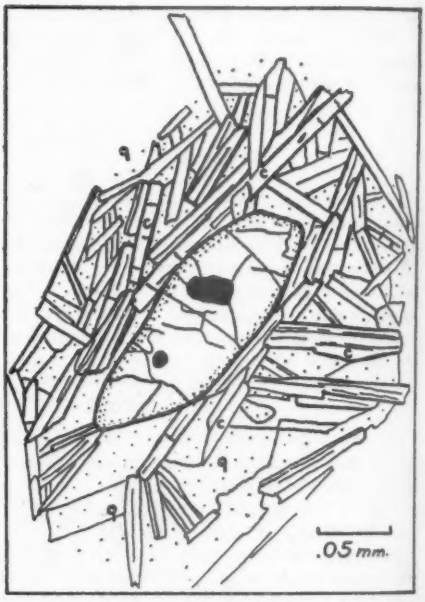



FIGURE 4. DETRITAL ZIRCON GRAIN IN SLATE
Grain is elongated parallel to schistosity and has tapered, somewhat frayed ends. Quartz (q) and chloritoid (c) also present.

lie parallel to the north-south or east-west crosshairs are in extinction. Furthermore, because of the positive elongation of the fibers, when a gypsum plate is inserted a rise and fall in colors is seen in the same quadrant as in the uniaxial positive interference figure.

Some of the mineralized cavities (Pl. 3, fig. 3) measure 0.72 x 0.35 mm. They may contain a rim of quartz crystals around a core of natrolite. The individual quartz crystals in the rim are 0.06 mm. long, and the natrolite center measures 0.32 x 0.18 mm. The delicate fibrous spherulites must have formed after deformation had entirely ceased.

Natrolite is a hydrous aluminum silicate which forms at low temperatures. According to Dana (1932, p. 654), it is one of the latest zeolites to crystallize. It could form from the alteration of albite by thermal waters according to the equation:

albite → natrolite

$$2NaAlSi_3O_8 + 2H_2O \rightarrow Na_2Al_2Si_3O_{10}\cdot 2H_2O + 3SiO_2$$

This equation also explains the presence of associated quartz.

### *Laumontite*

This zeolite is believed to be present in some late syntectonic quartz veins in slate. Observed in tiny pockets in a shattered grain of topaz, it was somewhat fibrous, white, length-slow, and had an extinction angle of about 30 degrees.

### *Topaz*

Many of the quartz veins in the slate and schist carry a little topaz. It occurs in ragged or irregular grains up to 5 mm. across. The grain boundaries indicate resorption by the surrounding quartz. The topaz crystals formed at slightly lower depths and were emplaced before the quartz in the vein had crystallized.

One crushed topaz grain suggests that the vein was late syntectonic.

### *Apatite*

Apatite is a rare accessory mineral in the rocks of the area. The grain size is variable, but 0.2 mm. is the maximum observed.

### *Garnet*

Minute porphyroblasts of garnet have been reported from several localities in the Peters Creek schist (Cloos and Hietanen, 1941, p. 130).

A garnet crystal 1.2 mm. in size was found in a calcite vein in the schist about 175 feet north of the railroad bridge over Peters Creek.

## Fossils

The writer found no fossils in the Peach Bottom slate during the two summers spent in the field. Several inorganic cryptic markings and concretions were found. The great shearing stresses to which the slates were subjected, the development of slaty cleavage, and the almost complete obliteration of bedding preclude the probability of preserved fossils. They might possibly survive in zones of less intense metamorphism, where slaty cleavage is parallel to bedding.

Old quarrymen in the district have told of seeing fern-like "fossils" in the slate, but none could be produced. Professor B. F. Howell informed the writer that several Peach Bottom fossils were in the collections at Princeton University. Professor Erling Dorf, of that institution, kindly lent them to the writer.

These fossils were found on a dump pile of one of the slate quarries near Delta in 1876. According to Percifor Frazer (1879), they were identified by Leo Lesquereux as *Buthotrephis flexuosa*. James Hall later examined the fossils and reported (Frazer, 1884): "... The conspicuous fossils have the character of marine vegetation, and the aspect of algae." The forms were replaced by films of iron pyrites; some are lobed, others branching. The edges are sharp, rather than gradational. These observations, plus the fact that one specimen showed considerable thickness, as if it might once have been a vesicular substance, suggested to Hall that the specimens were of organic origin. He referred two of them to *Buthotrephis* and considered their age to be "Hudson River" or "Quebec".

The writer verified the observations of Hall and submitted the fossils to several paleontologists. James Schopf (oral communication) thought the specimens were probably organic and resemble algae of the genus *B. newlini* David White, from the Upper Silurian Eurypterid beds at Kokomo, Indiana (David White, 1901). R. S. Bassler was confident that they were organic (oral communication). *Buthotrephis* is not a good diagnostic fossil. Its affinities, whether sea weed or algae, are uncertain. Nevertheless, it may be considered as a useful form genus. Bassler pointed out (oral communication) that similar specimens are well known from the Trenton and Cincinnatian limestones west of the crystalline rocks of the Appalachians. The bibliographic index of American Ordovician and Silurian fossils (Bassler, 1915) lists 20 forms assigned to *Buthotrephis*, ranging in age from Canadian? through Silurian.

Thus, the scanty fossil evidence in the Peach Bottom slate would suggest an Ordovician or Silurian age.

## Stratigraphic Relations

### *Age of the Slate*

James Hall was first to assign an age to the Peach Bottom slate when, on the basis of fossils he identified as *Buthotrephis*, he correlated the slate with the Ordovician "Hudson

River" shales (Frazer, 1884). Peter Lesley (1892, p. 140) thought that the slates might be Cambrian like those in New York and Vermont, but he had no evidence for this.

E. B. Mathews (1905) reasserted the "Hudson River" age for the slates, but admitted it was based on ". . . rather doubtful fossil evidence". Because he knew of no well defined and extensive separation between the Wissahickon schist and the slates, Mathews concluded that the Wissahickon was also Ordovician. In 1914 T. N. Dale (1914, p. 111) gave the age of the slates as Cambrian, but stated that no typical Cambrian fossils have been found in them.

Knopf and Jonas (1929) considered the slates pre-Cambrian because no unconformity could be demonstrated between the Cardiff conglomerate and the underlying "pre-Cambrian" Wissahickon schist.

J. H. Mackin (1935) reopened the question of the age of the Glenarm series and favored a Paleozoic age. At the same time, B. L. Miller (1935), working independently of Mackin, concluded that the Peach Bottom slates were Ordovician, and that a Precambrian age for the Glenarm series was highly improbable.

In 1939 Stose and Jonas (1939, p. 106) held the Peach Bottom slate and Cardiff conglomerate to be of probable Cincinnatian age and regarded the underlying Peters Creek schist as pre-Cincinnatian, contending that the Peach Bottom slate occurs in the same syncline as the Arvonia slate in Buckingham and Fluvanna Counties, Virginia. The two slates were tentatively correlated on a lithologic and stratigraphic basis (Stose and Stose, 1948).

### *Arvonia Slate*

In 1892, N. H. Darton found several fossils in the Arvonia slates and submitted them to C. D. Walcott, who identified the fossils as crinoids of Upper Ordovician age (Darton, 1892). Additional fossils discovered by S. L. Powell in 1909 were judged by R. S. Bassler to be middle Cincinnatian (Watson and Powell, 1911). In 1946, the same fossils were examined by C. A. Cooper, J. B. Reeside, Jr., and Edwin Kirk and a Maysville age was assigned to them (Stose and Stose, 1948).

If the Peach Bottom slates are equivalent to the Arvonia slates, they too would be Maysville. Stose and Stose believe the schist underlying the Arvonia slate is the same as the schist below the Peach Bottom slate. They have even extended the name Peters Creek to the schist in the Arvonia district (p. 405). No unconformity can be found between the Cardiff conglomerate and the Peters Creek schist in the Pennsylvania-Maryland area, but in the Arvonia slate district the thin basal conglomerate below the slate lies on the bevelled edges of the "Peters Creek" schist and the Columbia granite (Stose and Jonas, 1939, p. 104). On the basis of the unconformity above the "Peters Creek" schist in the Arvonia slate district, a time gap was postulated between the Cardiff conglomerate and the Peters Creek schist in the Peach Bottom syncline, where no unconformity can be demonstrated. The Peach Bottom slate and the Cardiff conglomerate were considered "of probable Ordovician age." and not to be included with the Peters Creek schist and the rest of the Glenarm series.

In 1946, Stose suggested that the rocks belonging to the Taconic sequence have been thrust over Martinsburg shales in eastern Pennsylvania, forming the Hamburg klippe and several smaller soles.

Two years later Stose and Stose (1948, p. 408) elevated the age of the Arvonia and Peach Bottom slates to Silurian or younger. Their reasons for this decision were as follows:

> "If the Arvonia slate is of Maysville age, as the faunal evidence has been interpreted, the slate must have been deposited before Taconic orogeny, in pre-Silurian time,. . . . If the Peters Creek formation, which is unconformably overlain by the Arvonia slate, belongs in the Taconic sequence, as has been suggested above, it is part of a sequence that ranges in age up to Lower Trenton. The time interval between the lower Trenton and Maysville is too short for the intrusion of the granite near Carys Brook (Arvonia slate district) and the subsequent uplift of the region and erosion deep enough to expose the granite; furthermore, during this interval elsewhere in the Appalachians, no such orogeny nor granitic intrusion are known."

It was then suggested that the Arvonia slate may not be Ordovician, but Silurian or younger, deposited after Taconic orogeny. It was further postulated that the Peach Bottom syncline extended 100 miles to the northeast and joined

the Green Pond Mountain syncline in northern New Jersey and southern New York, in which Silurian and Devonian sediments were deposited.

The present writer does not know of any basis for correlating the slate of the Peach Bottom syncline with the Silurian and Devonian sediments in the Green Pond Mountain syncline. Furthermore, it is difficult to see how the Arvonia slates can be Silurian or Devonian if they carry Maysville fossils.

The Ordovician rocks of Pennsylvania show a downward "migration" of the clay-limestone boundary from northwest to southeast. On the basis of recent fossil finds, the argillaceous Conestoga limestone in the Chester Valley area appears to be in part, of Beekmantown age (Swartz, 1948). Swartz considers it plausible that the Wissahickon schist may represent metamorphosed Beekmantown and Chazy sediments. In several places along the Steelton syncline, near Harrisburgh, the Beekmantown-Martinsburg contact appears to be unconformable. This suggests that emergence and erosion preceded the deposition of the Martinsburg (Swartz, 1948).

In the Arvonia slate district Stose and Jonas (1939) found an unconformity above the Peters Creek schist. If this unconformity also exists in the Peach Bottom syncline, an interval of erosion must have followed the deposition of the Peters Creek schist, of Beekmantown-Chazy (?) age. The Peach Bottom slate and the Cardiff conglomerate would then be post-Chazy, and could be Maysville, as are the Arvonia slates.

## Veins

Many thousands of white quartz veins and a lesser number of calcite veins are intruded into the rocks of the area. Pegmatitic veinlets occur in the schist south of the Peach Bottom syncline but are not very abundant.

The quartz veins range from 2 feet to a quarter of an inch wide, or less. The largest ones occur in fractured and brecciated zones in the slate. They are very irregular in form, pinching and swelling over short distances. Pyrite and chlorite, as well as minute amounts of muscovite and topaz, are present in the quartz veins. Some of the calcite veinlets carry quartz, biotite, and garnet.

The quartz veins show very little contact action, but sometimes chloritoid and chlorite are present along contacts (Pl. 3, fig. 4). Some of the veins in slate may be due to segregation of quartz during metamorphism of the argillaceous sediment. Quarrymen commonly remark that "flint" seams cleanse the rock and make the cleavage finer and truer (Mathews, 1898, p. 224).

The Peters Creek schist has in many places been soaked by quartzitic and pegmatitic solutions, as shown by epidotization of the schist adjacent to contacts, and by the larger size of porphyroblasts of albite and microcline in the vicinity of veins.

The veins are of several different ages. The oldest ones follow bedding and schistosity and show completely recrystallized quartz with granoblastic texture. Most of the recrystallized quartz grains, however, show strain due to subsequent deformation. The youngest veins transect all other structures and contain clear, unrecrystallized quartz. Most of the veins are syntectonic; they were introduced during the course of deformation.

Early veins in the Peters Creek schist are very irregular and may be folded or sheared (Fig. 5). Later ones form a network of large veins interconnected by smaller ones. In slate, some quartz veins ascended along the crests of steeply-plunging folds (Fig. 6 d). These are similar to the "saddle reefs" in the gold-bearing slates of Nova Scotia. Later veins filled transverse (*ac*) joint fissures which opened up when the rock was in tension parallel to *b*. In one instance (Fig. 6 c), the vein material appears to have moved along schistosity planes and then spread out along an *ac* joint.

A stereographic projection of 200 *c*-axes of needle quartz in a 3-inch quartz vein in slate is shown in Plate 11, diagram J 1. The quartz needles are about an eighth of an inch in diameter and grew normal to the sides of the vein. Small cavities in the vein could not have remained open if the vein were syntectonic. Neither is the vein entirely post-tectonic because the quartz is badly strained. It is probably very late syntectonic. The slate enclosing the vein is strongly lineated. The slate walls served

as a mold which imprinted the puckered lineations on the sides of the vein. Elsewhere, veins filling *ac* cross-joints transect *b*-lineations, indicating that the lineations are older than the veins.

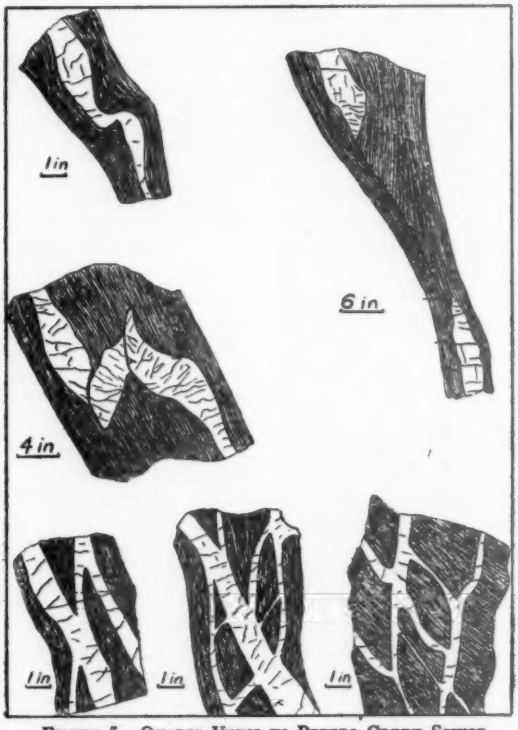



FIGURE 5.—QUARTZ VEINS IN PETERS CREEK SCHIST

## METAMORPHISM

### *Introduction*

Problems connected with metamorphic rocks can best be understood when the rocks are considered in terms of mineral facies as defined by Eskola (1920, p. 146).

Mineral facies change as metamorphism increases or decreases. But, as chemical reactions are frequently sluggish and proceed at extremely slow rates, mineral facies are ideally developed only when equilibrium conditions have been attained. Turner (1948 a, p. 93, 96) points out that synchronous deformation or intense hydrothermal activity will greatly speed up chemical adjustment of rocks in the lower grades of metamorphism and may also influence the composition of the chemical assemblage.

### *Peach Bottom Slate*

Over most of the area, the rocks belong to the Greenschist facies (Eskola, 1939, p. 360). Metamorphism increases in intensity outward from the slates at the center of the Peach Bottom syncline. In Dutchess County, New York, the muscovite slate facies is the lowest facies of the metamorphosed argillaceous rocks (Barth, 1936). The Peach Bottom slate of southeastern Pennsylvania and Maryland were likewise regarded as belonging to the muscovite slate facies of the Glenarm series (Cloos and Hietanen, 1941). Whereas chlorite is not present in the slates of Dutchess County, it is an im-

portant constituent of the Peach Bottom slates. Muscovite is not the only recrystallized mineral in these slates; chlorite and chloritoid have also been formed.

the slate were rich in alumina and iron. The amount of muscovite which could form depended on how much potash was available in the sediments. The presence of chlorite and

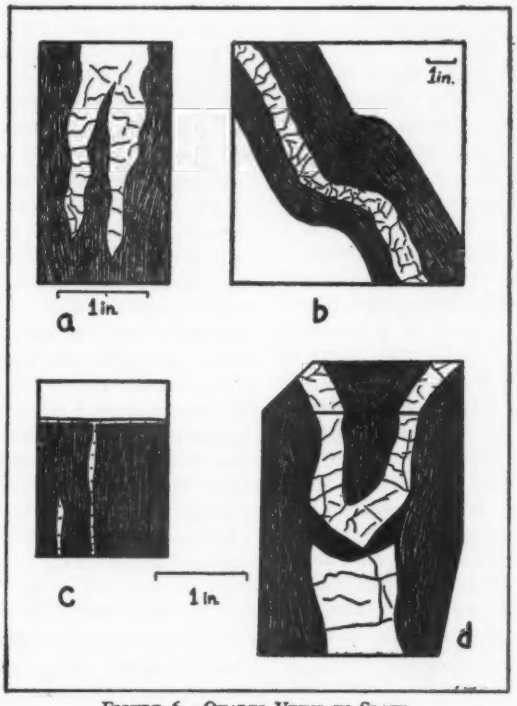



FIGURE 6.—QUARTZ VEINS IN SLATE

Chlorite and chloritoid are not as conspicuous as muscovite in thin sections, but X-ray analyses of two samples of Peach Bottom slate show an abundance of chlorite, the amount being approximately equal to that of muscovite. The slate inside the tight Peach Bottom syncline was metamorphosed beyond the stage represented by the muscovite slate facies, yet the rock is too fine-grained to be considered a schist. The Peach Bottom slates can therefore be regarded as constituting a muscovite-chlorite slate facies.

The degree of metamorphism of the slate is not uniform throughout the belt. It increases from southwest to northeast, in a general way, but it is locally higher in zones where shearing has been more intense.

The argillaceous sediments which gave rise to chloritoid indicate that potash was not present in sufficient quantity to combine with all the alumina and iron to form muscovite. Shearing stresses facilitated the formation of chloritoid, which is present towards the center of the syncline, in the slate and conglomerate. It is especially abundant in local zones of increased stress but is inconspicuous in the Peters Creek schist outside of the Cardiff conglomerate.

### *Cardiff Conglomerate*

The Cardiff conglomerate is a mildly feldspathic, somewhat argillaceous quartzite. Foliation is poorly developed, generally, but in places the rock has the aspect of a quartz schist. The effects of metamorphism were chiefly mechanical recrystallization and clastic deforma-

tion of quartz grains and pebbles. The argillaceous matter in the original sediment recrystallized to form muscovite, chlorite, and chloritoid.

### *Peters Creek Schist*

The Peters Creek schist shows lithologic differences and minor facies variation but it belongs essentially to Eskola's greenschist facies. Some of the quartzo-feldspathic beds in the Peters Creek schist are derived from arkosic sediments. Their feldspar content runs as high as 28 per cent. They resemble the rocks of subzone chlorite 2 and subzone chlorite 3 recognized by Turner in the Te Anau series of southern New Zealand (Turner, 1935, p. 345).

The Peters Creek schist is a muscovite-chlorite schist with a considerable quartz and albite content. Sericite predominates in some beds, and chlorite in others. Epidote and calcite are commonly present in minor amounts but they do not always represent original constituents of the rock. Calcite veinlets of igneous origin have epidotized the schist adjacent to the contacts. Very small amounts of chloritized biotite are not uncommon in the schist and the conglomerate, which indicates that the area has attained the biotite isograd. Many thin sections show small laths of biotite and brown pleochroic remnants of once-large flakes.

The formation of biotite may be retarded in quartzo-feldspathic rocks, as compared with pelitic schists (Turner, 1948 a, p. 96). Quartzo-feldspathic rocks can sustain higher shearing stress, thereby permitting extensions of the stability range of muscovite-chlorite into temperatures above those at which biotite first appears. Thus, the quartzo-feldspathic beds may have been subjected to a somewhat higher degree of metamorphism than their mineral composition would indicate.

### *Diaphthoresis*

The chlorite-muscovite schists are not to be regarded as diaphthoritic products of a biotite-muscovite schist. The evidence indicates that the biotite isograd had been attained in the progressive metamorphism of the rocks. On the other hand, there is not enough evidence to say that the Peters Creek schist has ever been metamorphosed to the stage of a biotite schist. These are not diaphthoritic chlorite schists. Diaphthoritic processes that have occurred were probably "... due to hydrothermal activity in connection with abundant pegmatitization" (Cloos and Hietanen, 1941, p. 135).

Tiny crystalloblasts of garnet are reported from several localities in the Peters Creek schist but their formation is subject to great local variation (Cloos and Hietanen, 1941, p. 130). Prophyroblastic albite is apt to be more conspicuous where the minute garnets occur.

The metamorphic rocks of the Appalachian Piedmont of Maryland and Pennsylvania do not show a uniform increase in grade of metamorphism from northwest to southeast; instead, there are local areas of more intense metamorphism. The Safe Harbor area is such a place of higher metamorphism within the chlorite schist facies. The locally higher grade of metamorphism is thought to be due to intense pegmatitization, perhaps indicating a magmatic intrusion below (Cloos and Hietanen, 1941, p. 136). Metamorphism is also much more intense around the gneiss domes north and west of Baltimore. Several of the domes show granitic cores (Broedel, 1937, p. 161).

### *Port Deposit Granodiorite Complex*

The Port Deposit granodiorite complex may have been the source of the pegmatites and veins which intrude the rocks of the Peach Bottom syncline. Emanations from this igneous body may also have given rise to the pneumatolytic minerals in these formations. The Port Deposit granodiorite lies 5 miles southeast of the Peach Bottom syncline and strikes roughly parallel to it. The igneous body intrudes the Glenarm series and other formations over an area 30 miles long and 9 miles wide (Hershey, 1937, p. 109).

## Megascopic Structures

### *Introduction*

Leith's usage of the terms flow cleavage and fracture cleavage (Leith, 1905; 1923) is followed in this discussion. In addition, planar structures are referred to as S-planes (Sander, 1930). The planes are designated $S_1$, $S_2$, $S_3$ ... in chronologic order, where possible. Bedding is $S_1$, flow cleavage is $S_2$, fracture cleavage is $S_3$ or

$S_4$. Strike readings are read from 0° to 180°. North is 0°, north 40° east is 40°, north 30° west is 150°. Plate 8 shows the statistical diagrams of the various structural features. These diagrams are equal-area net projections, as seen in the geographic position.

### *Bedding* ($S_1$)

Bedding has been superseded by schistosity as the dominant structure. Bedding can rarely be discerned in the slate, but it is more evident in the Peters Creek schist, especially where lithologic changes occur. Behre (1933, p. 368) recognized bedding in the slate only in about ten localities. The writer could add very few more.

The contact between the slate and the underlying Cardiff conglomerate is a bedding plane. Quartzitic intercalations and lenses in the slate also indicate bedding, which is generally parallel or subparallel to slaty cleavage. Shearing occurred more readily along bedding planes, as they were existing planes of mechanical weakness. Irregular joint-like "slaunts" suggest bedding, but they are considered dubious evidence (Behre, 1933, p. 368).

Stose and Jonas (1939, Pl. 19, 20) show four photographs of "probable" bedding. The writer saw these localities and does not think they show bedding. In the quartz schist south of the slate, along the west bank of the Susquehanna River, bedding dips 70° to the southeast and cleavage dips 55° to the southeast (Behre, 1933, p. 365). This indicates the south limb of a northward overturned syncline. Bedding is destroyed in the middle of the Peach Bottom syncline, but becomes more prominent outward.

Plate 4, Fig. 1 shows the contact between light-gray quartzitic schist and dark-green chlorite schist in a synclinal fold 180 feet north of the railroad bridge over Peters Creek. The bedding plane is very contorted and is followed by veinlets carrying quartz and calcite.

Thin quartzite beds are sometimes broken into lenticular "boudins" (Lohest, 1909). Figure 7 shows such a drawn-out bed in the Peters Creek schist. The gap between the boudins was filled by the flowing in of incompetent sericitic and chloritic layers. This illustrates stretching parallel to *a*.

In the Peters Creek schist, about half a mile from the slate belt, bedding is thrown into open folds which form large angles with h01 fracture cleavage.

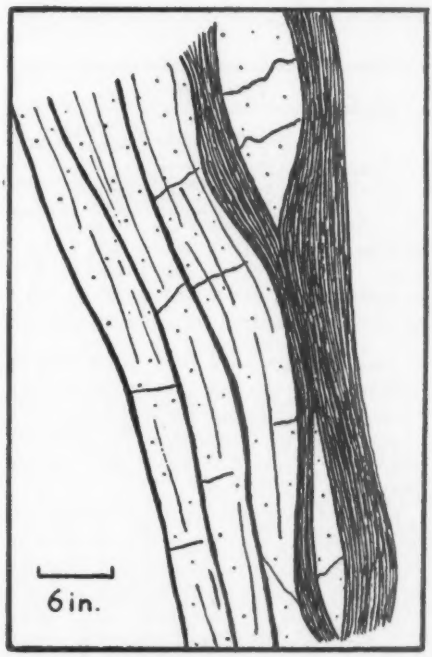



FIGURE 7. BOUDINAGE IN SANDY BED IN PETERS CREEK SCHIST

Argillaceous material has flowed in between the boudins.

### *Flow Cleavage*

Flow and fracture cleavage are usually distinguished. Flow cleavage is due to recrystallization and reorientation of platy minerals. Fracture cleavage is due to the formation of closely spaced planes of weakness or rupture. It also includes slip-cleavage.

Flow and fracture cleavages may result from shearing. Flow cleavage in argillaceous beds commonly passes into fracture cleavage in quartzitic beds. Cloos points out that there are all gradations between flow cleavage and fracture cleavage (Cloos and Hietanen, 1941, p. 24).

Flow cleavage (schistosity, slaty cleavage) is the most conspicuous structure in the slates.

In the Peters Creek schist, flow cleavage is well developed in sericitic and chloritic members. Quartz pebbles are flattened and elongated in the cleavage plane. Flow cleavage commonly parallels bedding, but in open folds the two planes may intersect at a large angle. Flow cleavage surfaces have been folded just as bedding. Flow cleavage usually shows lineations, which may be hair-fine or may appear as coarse puckerings. Coarse lineations are present where flow cleavage is folded and fracture cleavage is well developed.

TABLE 3. STRIKE OF FLOW CLEAVAGE IN FORMATIONS

| Formation | Lancaster County | | York and Harford Counties | |
|---|---|---|---|---|
| | Average strike | Number of readings | Average strike | Number of readings |
| Pb slate..... | 51° | 197 | 62° | 228 |
| Ccgl..... | 51° | 30 | 68° | 37 |
| Pc schist..... | 57° | 282 | 55° | 105 |

Plate 8, A–F, shows stereographic projections of flow cleavage orientations. The mean strike of flow cleavage in the formations is indicated in Table 3. It is more northerly in the Lancaster County part of the area. Dips usually range from vertical to 65° to the southeast, but steep northwest dips and gentle dips are also present.

### *Folds*

None of the several types of subsidiary folds within the Peach Bottom syncline are very large. Most folds in the Peters Creek schist are undulatory and open and represent folded bedding. Within a half mile of the center of the syncline, folds become more isoclinal and slightly overturned to the northwest. Folded bedding is easily distinguished in sandly layers, but in argillaceous beds it is obscured by strongly developed flow cleavage. Several folds in the Peters Creek schist are illustrated in Figure 8.

The Cardiff conglomerate is a hard, competent formation displaying no folds in outcrops. However, flexure folds may be present in the conglomerate underneath the slate, towards the interior of the syncline.

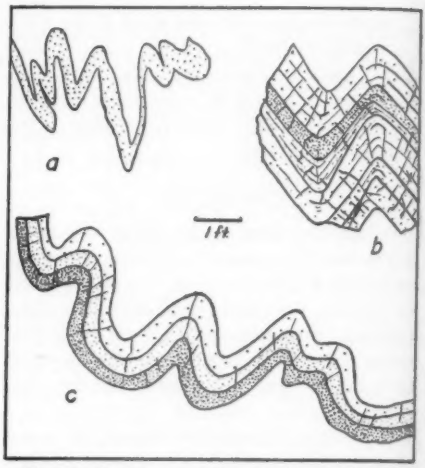



FIGURE 8. FOLDS IN PETERS CREEK SCHIST

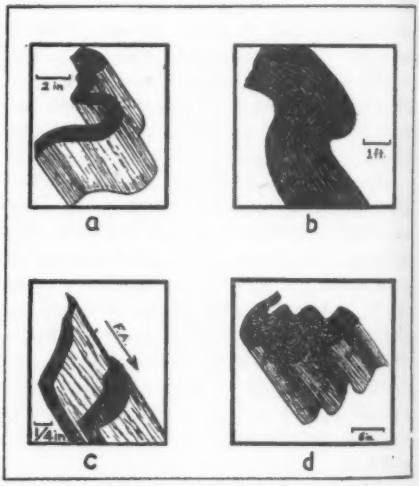



FIGURE 9. FOLDS IN SLATE

According to Stose and Jonas (1939, p. 100), the slate is folded into three minor synclines and two minor anticlines, with a gentle northeast pitch. Several exposures in the area faintly suggest isoclincal folds, but the writer was unable to verify their structural picture.

Figure 9 shows several types of folds observed in the slate. Drag folds formed by slip along foliation planes are common. The folds have small amplitude and extend for very short distances along the fold axis. The folds show

sharp and often intricate bending of slaty cleavage planes. The wave length of these folds is seldom more than 1 foot, and is commonly only about 2 inches. The folds may affect slaty cleavage to a slight distance, sometimes only

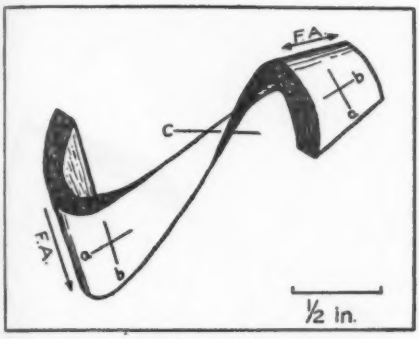



FIGURE 10. FOLDS IN SLATE SHOWING ROTATION OF *a* AND *b* AXES ABOUT *c*

1 inch, whereupon they grade into weak crinkles or lineations.

Figure 10 shows two small adjacent folds in the slate along the east bank of the Susquehanna River, about 2150 feet northwest of the railroad bridge over Peters Creek. These folds clearly indicate the rotation of the *a* and *b* axes about the *c* axis. Ten feet south, a sharp fold 2 feet across plunges 10° to the NNE. It grades rapidly into a more open flexure which plunges 50° to the ESE.

Another specimen of folded slate from the east bank of the Susquehanna River is shown in Figure 11. The fold axis, *b*, strikes ENE and is slightly flexed about an axis $b'$ (= *a*). This is an example of a cross fold. According to Fairbairn (1942, p. 78), the magnitude of a cross or saddle fold may be very minute but it is permissive evidence of longitudinal shortening, supplementing the evidence of *ac* tension fractures.

### *Fracture Cleavage* ($S_3$)

*General statement.*—Fracture cleavage (false cleavage, slip cleavage) transects flow cleavage and bedding in all formations, but is strongest in the slate (Pl. 7). Fracture cleavage approximately parallels axial planes of folds and crenulations.

In many places, flow cleavage planes are displaced by slip on fracture cleavage. The displacement ranges from less than a quarter of a millimeter to more than 6 millimeters. Larger displacements result in slip faults. Several narrow zones in the slate and schist show strong, closely spaced fracture cleavage. Shearing along

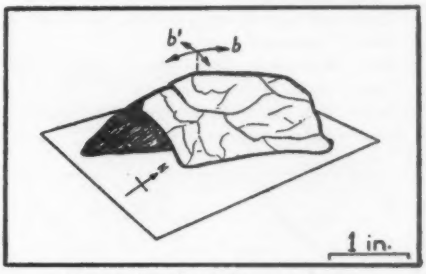



FIGURE 11. CROSS FOLD IN SLATE

these planes has been so intense that earlier flow cleavage is almost obliterated.

Movement of solutions is facilitated by fracture cleavage planes. Muscovite and other minerals later crystallize along the planes and emphasize their anisotrophy. Sander (1930, p. 172) called this *Abbildungskristallisation*, which was translated by Suess (1931, p. 78) as *mimetic crystallization*. Slate slivers or pencils form where fracture cleavage intersects slaty cleavage at a high angle.

*Orientation of fracture cleavage.*—The strike of fracture cleavage in the slate west of the Susquehanna River varies considerably. Along the river the strike is generally southeast, and the dip is steep to the northeast. Elsewhere, west of the river, the strike is a little north of east, and the dip is steeply northwest. In Kell quarry, fracture cleavage strikes north and dips steeply west. The orientation of fracture cleavage in Lancaster County is much more uniform. The average of over 200 readings of fracture cleavage in the schist and slate in Lancaster County indicates a strike of 119° and a dip of 77° to the northeast. The angle between the strike of schistosity and fracture cleavage is nearly always 65°.

*Two fracture cleavages.*—Two sets of fracture cleavage, one horizontal or gently dipping, and the other steeply dipping are present in these localities:

1. Road cut 0.4 mi. east of Susquehanna River, 30 ft. northwest of slate-conglomerate contact.
2. East shore Susquehanna River, 300 ft. northwest of slate-conglomerate contact.
3. East shore Susquehanna River, 400 ft. northwest of slate-conglomerate contact.
4. Near the center of Gorsuch quarry, east shore Susquehanna River.
5. Creek bed east of road, one mi. northeast of Greene, Lancaster County.
6. West shore Susquehanna River, 50 ft. northwest of slate-conglomerate contact.
7. West shore Susquehanna River, 300 ft. northwest of slate-conglomerate contact.

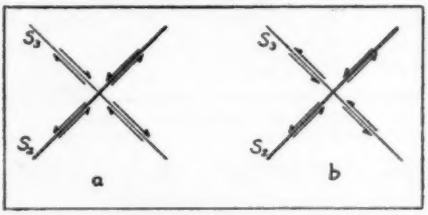



FIGURE 12. RELATIVE MOVEMENT ALONG SLATY CLEAVAGE ($S_2$) AND FRACTURE CLEAVAGE ($S_3$)

Steeply dipping fracture cleavage is much more abundant than horizontal fracture cleavage. Where both are present, either one may be more prominent. The stronger set may become almost inconspicuous within 1 or 2 feet.

Some of the steeply dipping fracture cleavages appear to be younger than horizontal ones. In other instances, the reverse is true. This observation confirms Billings' statement (1947, p. 128) that where two sets of shear fractures are present, individual fractures of one set may be younger or older than fractures of the other set, but there is no systematic age difference.

In the horizontal set or gently dipping fracture cleavages, the relative movement of the "hanging wall" is uniformly to the northwest. In the steep fracture cleavages, however, the movement is not always in the same direction. The laminae northeast of the steep fracture cleavage planes usually show a relative movement to the southeast (Fig. 12 a), but in some p aces the movement is to the northwest (Fig. 12 b).

Figure 12, showing relative movement along intersecting shear planes, is presented in Sander (1934, Fig. 1) and in Ingerson (1936). Figure 12 indicates that the two slip planes are related to one act of deformation, possibly with internal rotation about B. Figure 12 b indicates that the slip planes are due either to two unrelated deformations or to a single deformation accompanied by external rotation not necessarily about a B-axis (Ingerson, 1936, p. 170).

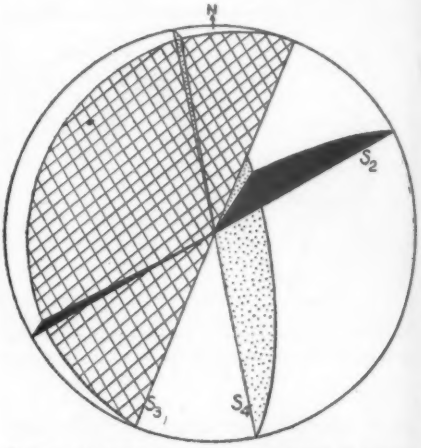



FIGURE 13. STEREOGRAPHIC PROJECTION IN LOWER HEMISPHERE OF SLATY CLEAVAGE ($S_2$) AND TWO FRACTURE CLEAVAGES ($S_3$ AND $S_4$)

Figure 13 is a stereographic projection of slaty cleavage ($S_2$) and two fracture cleavages ($S_3$ and $S_4$) in a specimen of slate. One fracture cleavage dips steeply to the east-northeast, the other gently to the south. The projection does not readily indicate that the fracture cleavages are symmetrically arranged with respect to slaty cleavage. However, this is immediately apparent if the fracture cleavages are projected into the plane of slaty cleavage (Fig. 14) by rotating the pole of slaty cleavage to the center and rotating the other planes accordingly. Similar symmetrical relationships of the two fracture cleavages with respect to flow cleavage are shown wherever the three cleavage planes are present.

An early attempt to explain two sets of fracture cleavage in slate was made by H. B. Muff, who observed two sets of strain-slip-cleavage in the Craignish phyllites of Scotland (Peach, Kynaston, and Muff, 1909). One of the cleavage planes is vertical, the other is horizontal or gently dipping towards the west. Both

cleavages are inclined 45° to the slaty cleavage. According to Mead (1940, p. 111), they would then be oriented approximately 45° from the axis of the compressive resultant of force. In Muff's diagram (Peach, Kynaston, and Muff,

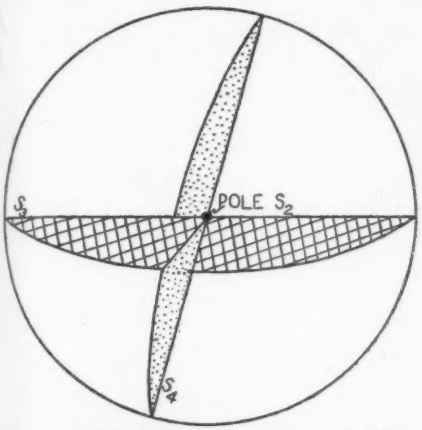



FIGURE 14. SAME AS FIGURE 13, ROTATED IN PLANE OF SLATY CLEAVAGE
Pole of $S_2$ is at center.

Fig. 3), the three cleavages appear to intersect in the fold axis. Muff states that

"... the strain-slip cleavage, the cleavage, and the folding are recognized as concomitant, though not simultaneous, effects of the same lateral pressure, and do not necessitate two or more distinct periods of earth-stress in the district. It should be pointed out, however, that a slight veering of the direction of maximum pressure towards the close of the period of earth-stress is required to account for the slight difference in strike between the strain-slip cleavage and the other structures."

According to Turner, (1948 b, p. 561), where several sets of active slip surfaces are fairly symmetrically arranged with respect to the AB plane of the ellipsoid, the compressive force was essentially normal to schistosity.

Unlike the example cited by Muff, the two fracture cleavages in this area usually do not intersect in the fold axis. Their line of intersection may be at a large angle to the slaty cleavage (*ab* plane), and inclined 10° or more to the *c* axis. The line of intersection of fracture cleavages is not necessarily a fabric axis of the rock.

The near-horizontal fracture cleavage intersects slaty cleavage in B, while the steeply dipping fracture cleavage intersects slaty cleavage in B′. B ∧ B′ is 73°. In the specimen illustrated in Figure 15, B is usually strongly overprinted by B′, but this relation is not consistent in other specimens. The two axes, B and B′, are largely contemporaneous, having formed during the same period of deformation, but rotation about B began slightly before rotation about B′.

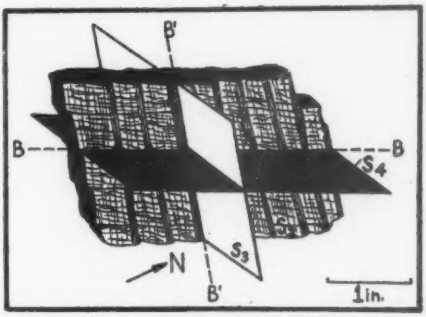



FIGURE 15. SLATE SHOWING TWO FRACTURE CLEAVAGES
($S_3$ and $S_4$) in slate.

In Figure 16 a three S-planes are easily distinguished. $S_2$ is flow cleavage, and represents the *ab* plane. $S_3$ and $S_4$ are fracture cleavages. Muscovite flakes have been sheared into $S_3$ without breaking, but those flakes which were bent into $S_4$ are broken. The directions of relative movement along the S-planes are indicated in Figure 16 b.

*Hypotheses on multiple-fracture cleavage.*—Becker's strain hypothesis would have attributed the unequal development of $S_3$ and $S_4$ to different angular velocities of the shearing planes as they rotated through the rock (Becker, 1893). The shearing plane which moves at the lower angular velocity ($S_3$) permits mineral grains enough time to flow into the shearing direction. On the other hand, the mineral particles are moved too rapidly into the plane which sweeps through a large arc ($S_4$), and consequently are bent and fractured. This hypothesis, though ingenious, has been shown to be untenable (Fairbairn, 1942, p. 92).

Schmidt's strain hypothesis states that rupture and gliding planes are subparallel to direc-

tions of maximum shearing strain ($S_3'$ and $S_4'$), which lie 45° between the major and minor strain axes (Schmidt, 1932). Furthermore, although the direction of the strain axes may change, the position of the shear planes remains unaltered.

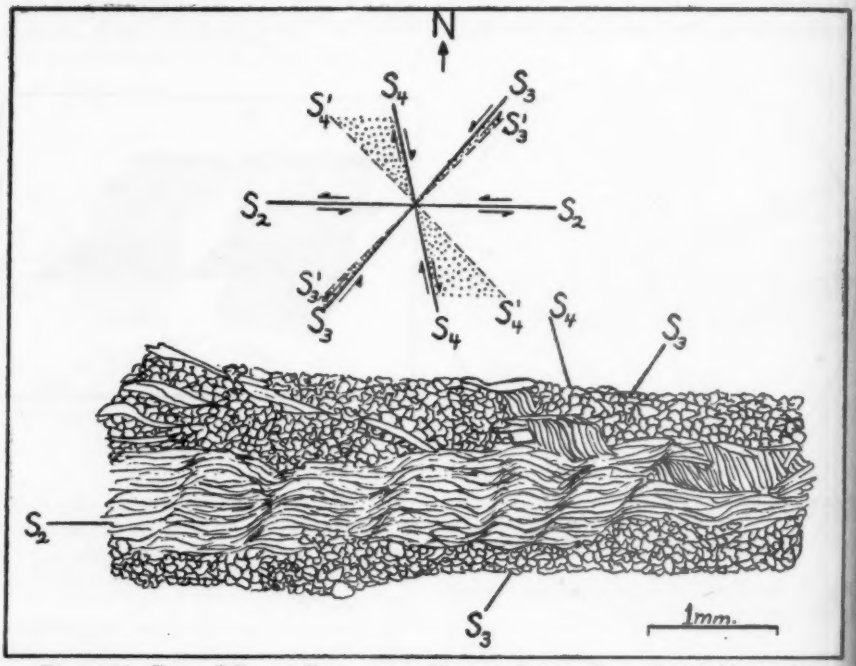



FIGURE 16.—THREE S-PLANES FOLLOWED BY MICAS IN PETERS CREEK SCHIST (a) AND (b) RELATIVE MOVEMENT ALONG S-PLANES

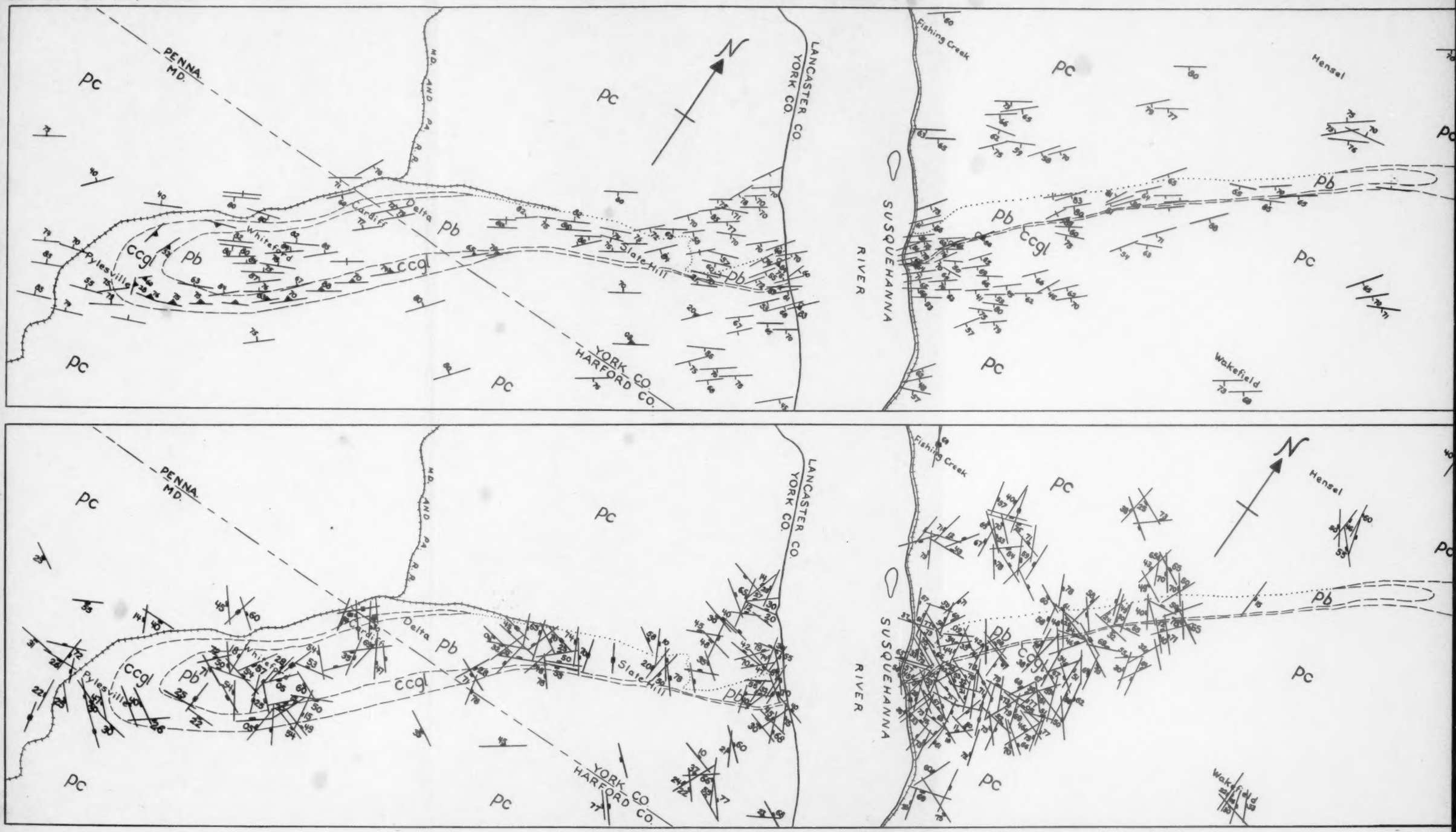
The writer agrees that axes of strain rotate, but he is of the opinion that fracture cleavage or shear planes in incompetent rocks such as slate and schist also undergo rotation. Due to internal friction, the rotation of glide planes probably lags behind that of strain axes. Thus, Schmidt's hypothesis of strain cannot be used in its entirety as an explanation of the relationship and character of the glide planes in Figure 25 a. According to Schmidt, the better of two glide surfaces lies more nearly parallel to a line of maximum shearing strain. The mica planes along $S_3$ (Fig. 16) are not broken. $S_3$ is a fairly good shear plane, and lies very close to one of Schmidt's theoretical planes of maximum shear, $S_3'$, (not considering rotation). But S which is a poor glide plane and contains bent and broken mica flakes, is removed about 32° from the original position of the second theoretical plane of maximum shear ($S_4'$).

*Fold Axes*

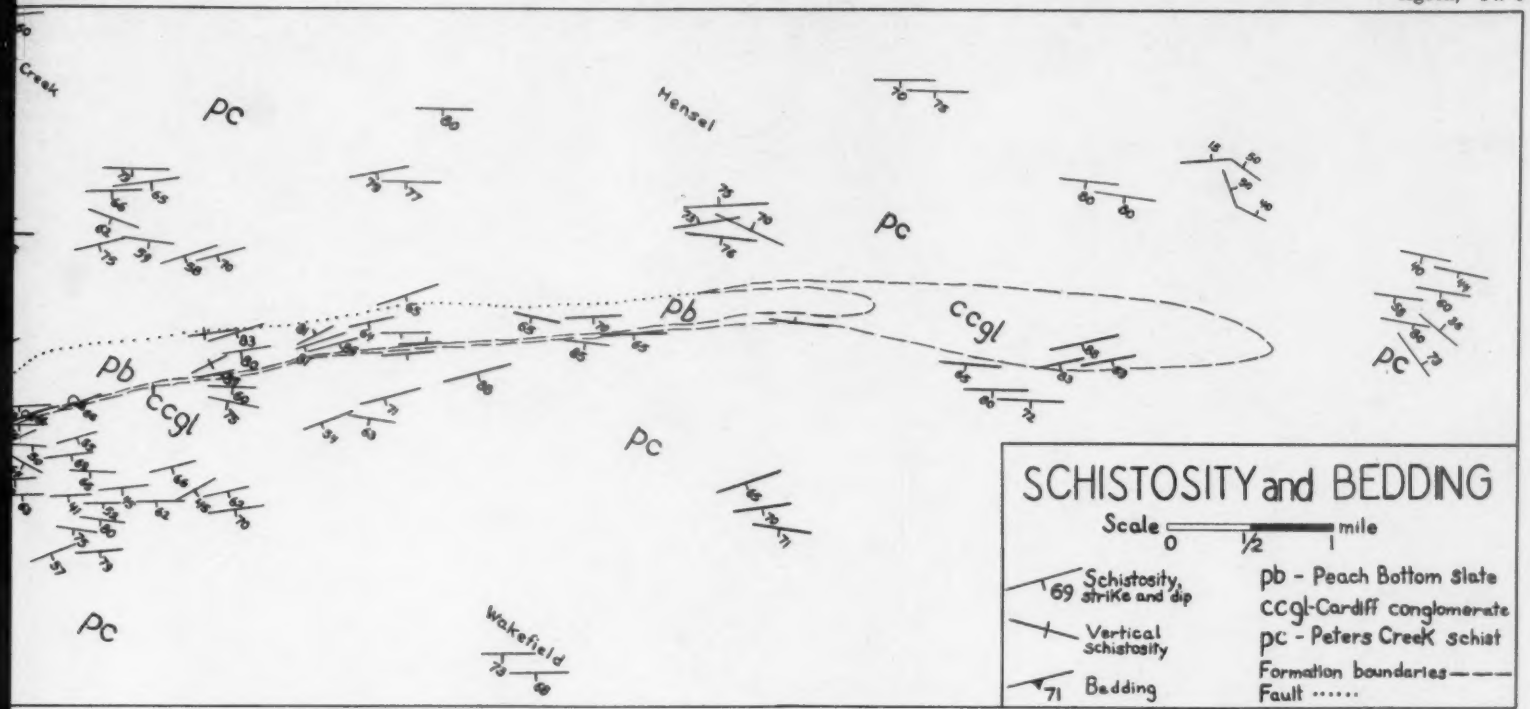
Fold axes can be measured directly in folds and they can also be determined by measuring the intersection of flow cleavage or bedding and fracture cleavage. These planes intersect in a fold axis, *b* or *b'*. The intersection is commonly marked by a lineation.

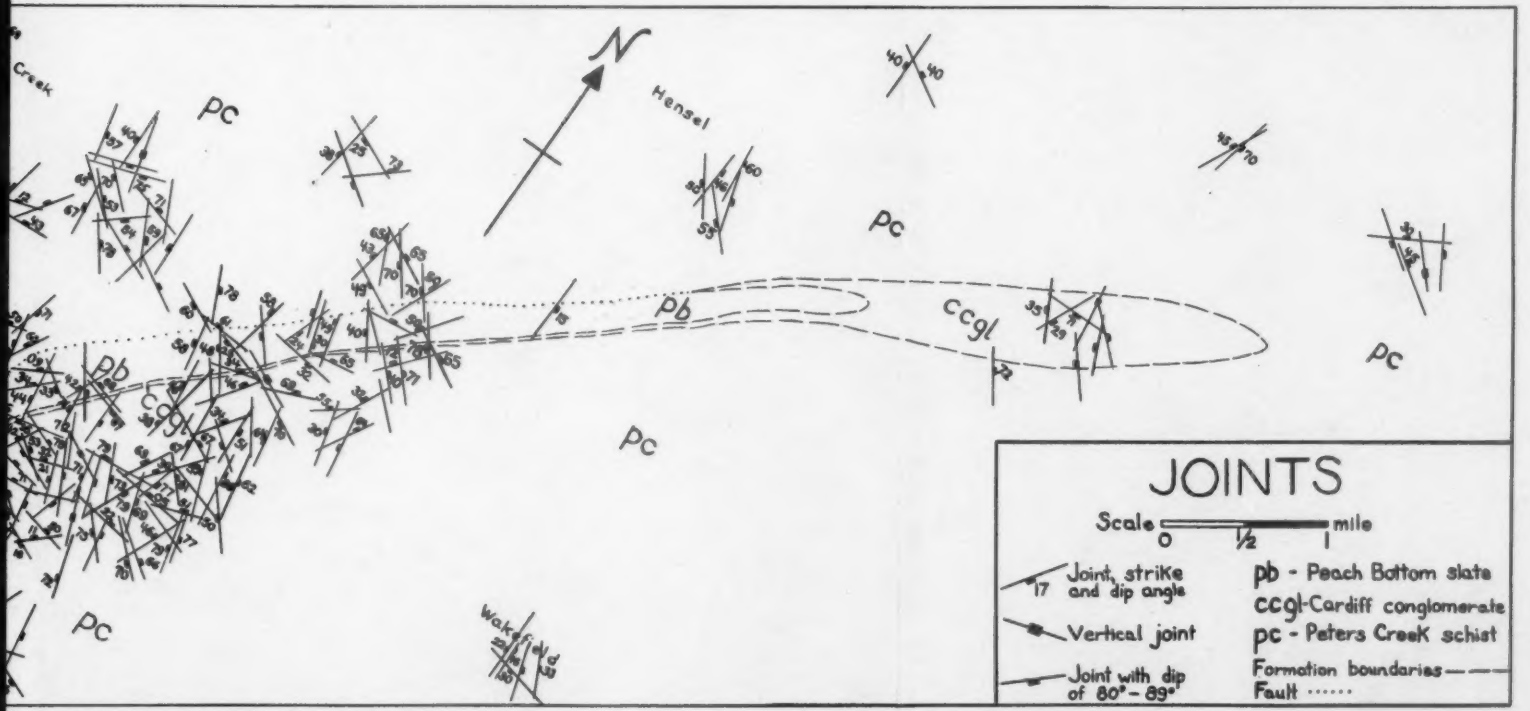
Fold axes do not show a uniform orientation in this area (pl. 7) as they do a few miles north, where they are almost horizontal and plunges exceeding 45° are rare (Cloos and Hietanen, 1941, p. 22). In the vicinity of the Peach Bottom syncline, however, near-horizontal fold axes are much less common than steep ones.

Stereographic projections of 453 fold axes and lineations (Pl. 8, M, N, O, P, especially O) show that the fold axes (and lineations) are spread out along the trace of flow cleavage. Most of the fold axes east of the Susquehanna




REGIONAL MAPS OF SCHISTOSITY, BEDDING, AND JOINTS




Y, BEDDING, AND JOINTS

River plunge 50° to 60° east, few are horizontal.

Some of the fold axes in the Peters Creek schist west of the Susquehanna River plunge 60°–70° east, but a greater number plunge more gently northeast (Pl. 8, P). Fold axes in the slate west of the Susquehanna River strike in various directions and plunge from horizontal to vertical (Pl. 8, O). Numerous readings of vertical fold axes and lineations were taken, as indicated by the 13 per cent maximum in Plate 8, O. Almost as many fold axes plunge 50° to the southwest as do 50° to the northeast. Northeast-plunging fold axes are more numerous than those plunging southwest.

Distribution of fold axes along the trace of flow cleavage (schistosity) reflects the rotation of the *b* axes about *c*. The amount of rotation in various parts of the area ranges from 0° to 90°. There has been more rotation in the slate than in the schist, especially west of the Susquehanna River.

A schematic longitudinal section of the slate in the Peach Bottom syncline is presented in Plate 7. The lines do not indicate beds, but rather directions of plunge of fold axes and lineations in different parts of the syncline. Rotation of fold axes about *c*, with the development of cross folds, is especially marked in several transverse zones.

### *Lineations*

The Peach Bottom slate and Peters Creek schist commonly show lineations parallel to *b*. They range from very faint striations to coarse, ridge-like puckerings. The Cardiff conglomerate and quartzitic beds in the schist are very rarely lineated, but where micaceous films or layers are present, faint striations can sometimes be seen.

The lineations are due to crinkling of flow cleavage surfaces, or intersections of flow cleavage with bedding or fracture cleavage (Pl. 5). Most often they are due to intersection of flow cleavage and fracture cleavage.

The strike of lineations is generally between northeast and southeast, and the plunge ranges from horizontal to vertical. Some lineations plunge westward. Plate 8, M, N, O, P are stereographic projections of 453 fold axes and lineations, the discussion of these diagrams previously given under "fold axes" applies equally well here.

In a few localities, two sets of lineations, one sub-horizontal and the other steeply plunging intersect at an angle of about 73° (Pl. 5, fig. 3), but the angle may vary considerably. Either the horizontal or the steep lineations may appear to be older, but there is no regional or even local uniformity. Either set may be more prominent; however, the steep lineations are usually finer and more closely spaced than the horizontal ones. Three sets of lineations are shown in Figure 4 of Plate 5.

### *Joints*

A regional joint map is presented in Plate 6. Strike, cross, and oblique joints are present but cross (*ac*) joints are much more numerous than others. The joint patterns are different on the east and west sides of the Susquehanna River. However, they are similar in all formations on each side of the river. Plate 8, G, H, I, J, K, L, are stereographic projections of the 1345 joints measured in the area. Most of the joint maxima (poles) lie in or close to the flow cleavage plane, indicating that there are many *ac* joints lying approximately normal to fold axes with various degrees of plunge. The cross joints are not quite normal to fold axes, but are at an angle of about 80°.

J, K, and L of Plate 8 are stereographic projections, in equal area nets, of the joints west of the Susquehanna River. The joint normals are largely concentrated at the periphery and near the center of the projections, indicating that the joints are, for the most part, either steeply dipping or flat-lying. Most of the cross joints strike to the northwest and stand vertical, or dip steeply to the northeast or southwest. A few steeply dipping longitudinal and oblique joints strike north or northeast.

The concentration of joint poles near the center of the projection indicates a large number of flat-lying joints, some of which are cross joints of vertical or steeply plunging folds. The joint maxima at the center of the diagram correspond to the central maximum for fold axes and lineations in the slate (Pl. 8, O). The extensions of the central joint maximum towards the north and south are due to gently dipping oblique and longitudinal joints.

In Lancaster County, most of the joints strike northwest and dip moderately to steeply towards the northeast or southwest. These are cross joints related to the regional *b* axis, which

strikes northeast. Cross joints approximately perpendicular to $b'$ are also common. They trend northeast and dip northwest or southeast, usually at a more gentle angle than the cross joints which strike northwest. The northeast-

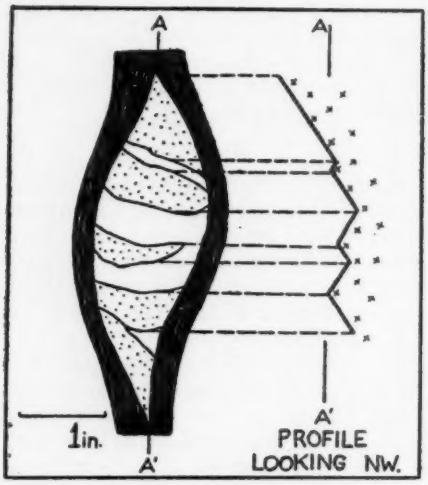



FIGURE 17. TWO SETS OF TRANSVERSE JOINTS IN QUARTZ VEIN

trending joints have a smaller dip angle because they complement $b'$, which plunges steeply towards the southeast. Joint maxima in the schist and slate (Pl. 8, G, I) lie 5° to 20° from the maxima for fold axes and lineations (Pl. 12, M, N), illustrating the approximately normal relationship between fold axes and cross joints.

The strong maximum in the southwest quadrant of the joint diagram for the Cardiff conglomerate (Pl. 8, H) indicates that many of the joints in the conglomerate strike northwest and dip steeply to the northeast. These joints appear in the same position as fracture cleavage in the slate and schist in Lancaster County, and they are probably analagous to fracture cleavage. However, in the Cardiff conglomerate, as in other competent formations, these joints show no shear displacement, as is shown along fracture cleavage planes in the slate and schist. The conglomerate frequently breaks into rhomboid blocks whose sides are $S_1$-$S_2$ planes, and whose ends are formed by two sets of joints intersecting in an angle of 60°. One of these sets of joints commonly strikes northwest and dips steeply northeast. The other set are BC joints which dip gently to the southeast.

Quartz veins may be well jointed. Figure 17 shows a quartz vein with two sets of transverse joints. Both strike about 142° and are inclined 62° to each other, suggesting shear joints. Figure 17 shows that the joints which dip 56° to the northeast are larger and smoother than those which dip 62° to the southwest.

Almost horizontal "big flat" joints can be seen in several of the slate quarries in York County. These somewhat undulatory shattered zones are spaced more closely towards the top of the slate. Water percolating along the joints weathers the slate and forms a red clay which is, in places, more than 2 inches thick.

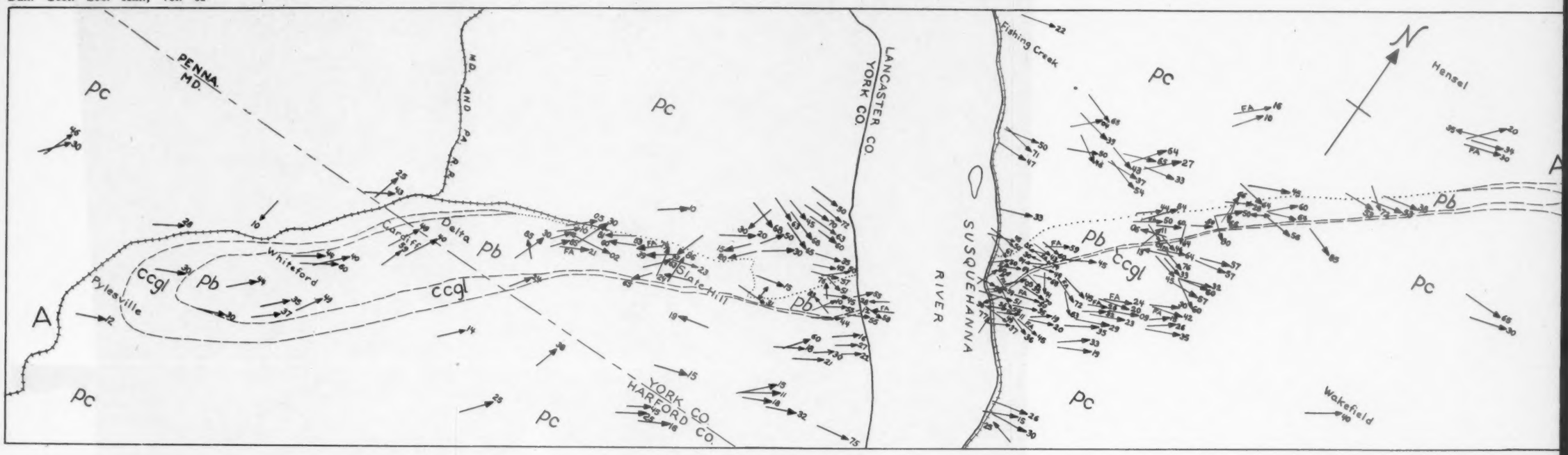
All joints formed at a late stage of deformation. A few late quartz and calcite veinlets filled joints which were open at the time the solutions moved through the rocks. Figure 6 c shows a quartz veinlet which ascended along slaty cleavage planes and then spread out along a cross joint. Striae on calcite- and quartz-coated joints indicate movement. Some striated vein quartz was seen on dumps, as at the Gorsuch quarry.

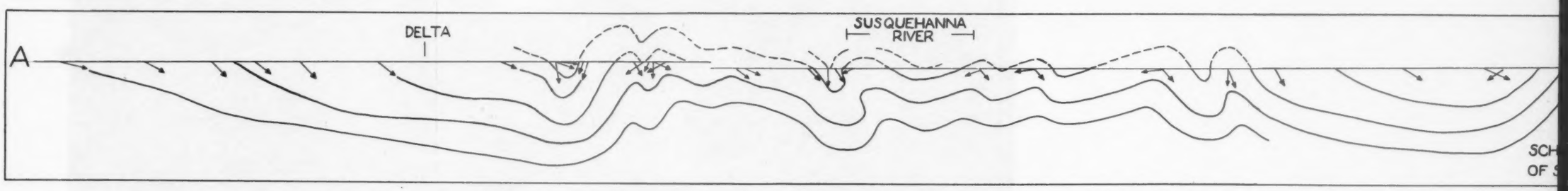
### *Grain*

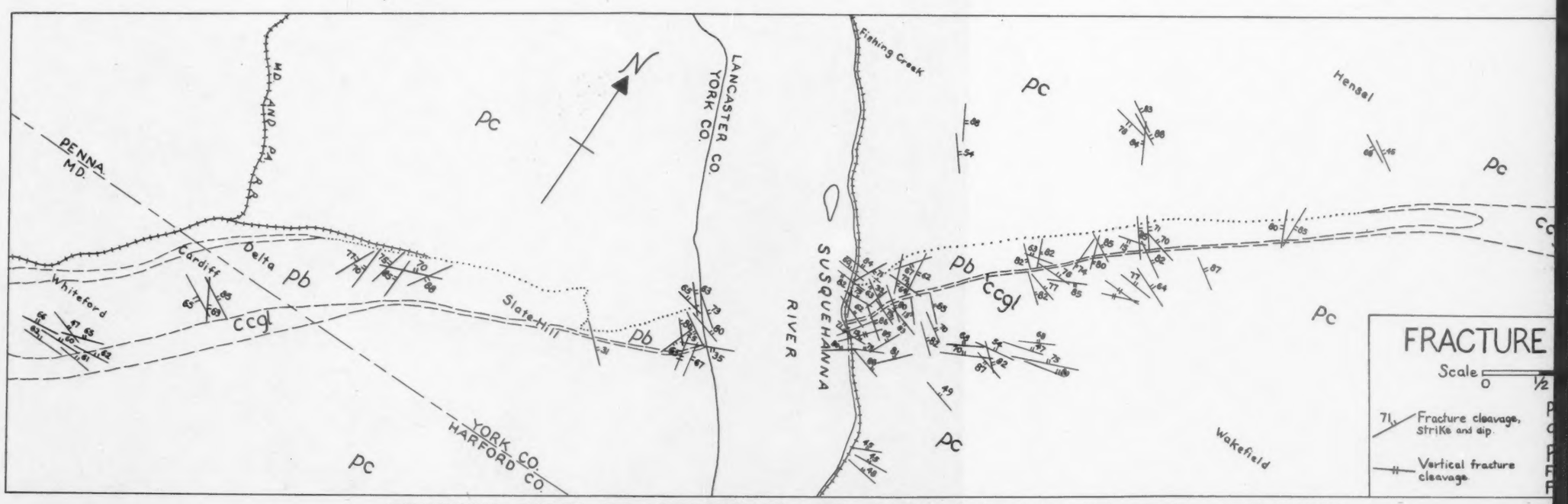
Grain has been defined as "... a second direction of ready fracture far less marked than that of the cleavage" (Behre, 1933, p. 31). The grain in slate is parallel to the longest dimension of mica, chlorite, and other minerals. According to Dale (1914, p. 41) grain may be indicated by an obscure striation on slaty cleavage, in a direction nearly parallel to the dip of cleavage. He also notes that in some localities there is hardly any grain.

Behre observed the "obscure striations" referred to by Dale, on many Pennsylvania slates. If these are the same as the ones discussed in this paper, then in the Peach Bottom slate, at least, the striations are due to the intersection of fracture cleavage with slaty cleavage. The striations or lineations are parallel to $b$ or $b'$.

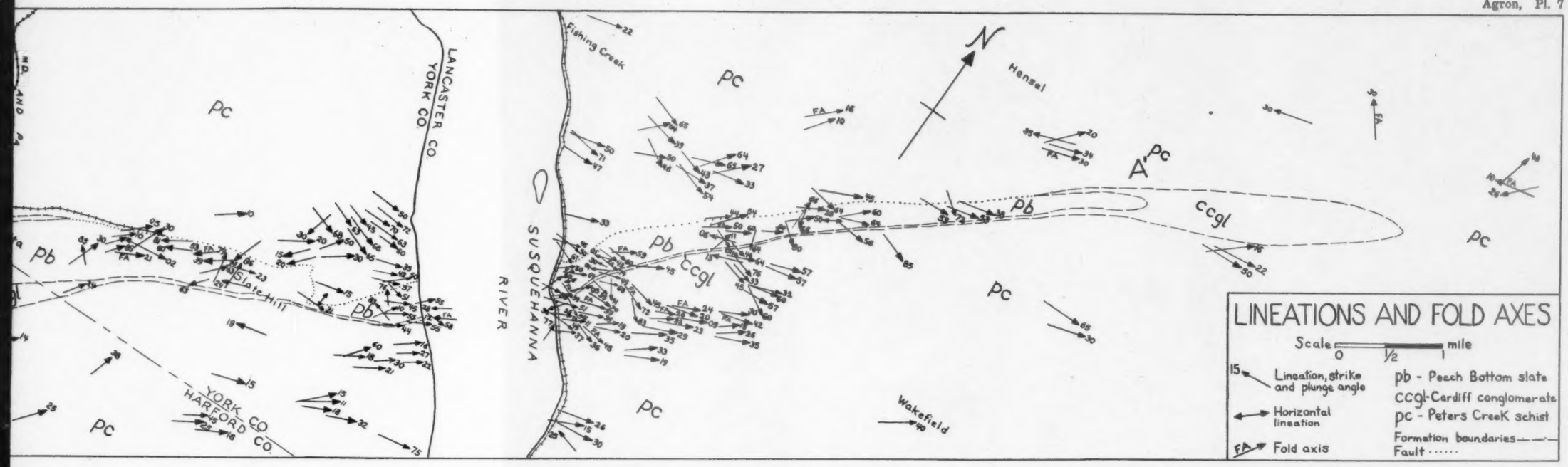
According to Behre (1933, p. 34, 372), the grain in the Peach Bottom slate, as elsewhere in Pennsylvania, stands nearly vertical and trends about N45°W, approximately at right angles to the slaty cleavage. This is the $ac$ direction with respect to $b$ and the regional tectonic strike. It

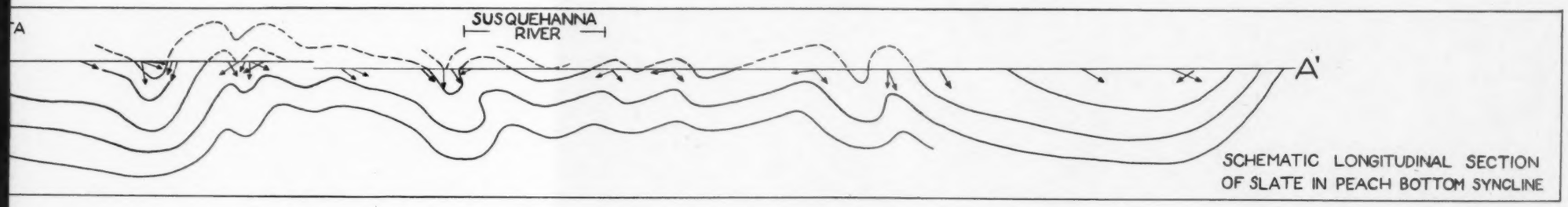


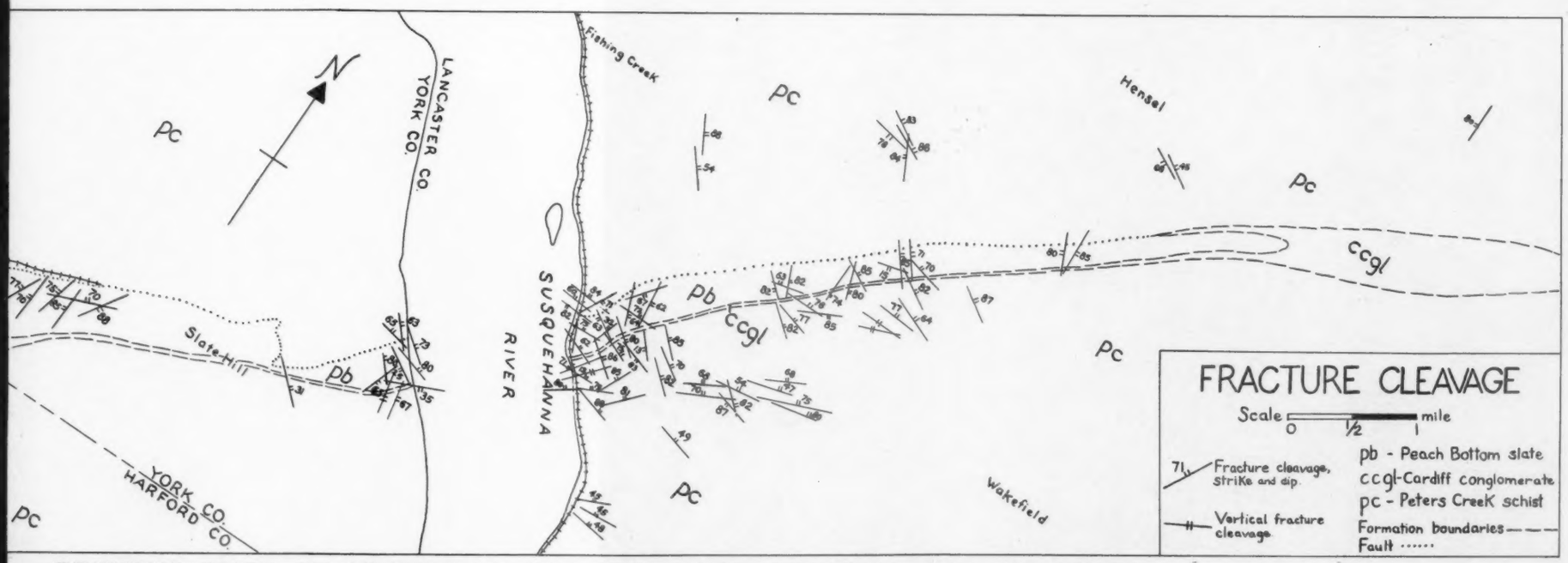




REGIONAL MAPS OF LINEATIONS AND FOLD AXES AND FRACTURE CLEAVAGE




**REGIONAL MAPS OF LINEATIONS AND FOLD AXES AND FRACTURE CLEAVAGE**

is also the fracture cleavage direction for *b'*, which plunges southeast. The *ac* or grain direction shows local variation of dip. In the southwest part of the syncline, the grain dips to the southwest. Further northeast, it is horizontal in places, while in the Gorsuch quarry, on the east bank of the Susquehanna River, it dips steeply to the northeast. This circumstance may have been overlooked when the Gorsuch cut was started and the slate worked eastward from the river. Because the grains dips northeast, it was more difficult and expensive to remove blocks of slate from the west than it would have been from the east. More efficient and economical operation would have resulted if the quarry had been opened further to the northeast and worked westward towards the river.

### *Synoptic Diagrams of Structures*

The synoptic diagrams (Pl. 8, S, T, U, V, W) show the nearly constant relationship between schistosity, fracture cleavage, and the major joint direction in the three formations east of the Susquehanna River. The lineations are located at the intersection of schistosity and fracture cleavage, approximately at the pole of the cross joints.

### *Faults and Thrusts*

*Introduction.*—Faults and thrusts, usually of small displacement, are not uncommon in the area. Small thrusts towards the northwest can be seen in many outcrops in the slate and schist. Along any one plane the displacement may be only a fraction of an inch, but it becomes significant when the aggregate of may thrust surfaces is considered. A lateral displacement of 1½ inches per foot along parallel fracture cleavage planes is shown in Figure 5 of Plate 4.

A small displacement towards the northwest in sandy and argillaceous beds in the Peters Creek schist is shown in Figure 18.

*Low- and high-angle thrust fault in schist.*—A low- and a high-angle thrust fault, indicating lateral compression in the Peters Creek schist, can be seen along the railroad cut 1350 feet northwest of Peters Creek (Pl. 4, fig. 2). This fault somewhat resembles Behre's sketch of a bedding slip fault in one of the Slatington, Pennsylvania quarries (Behre, 1933, Fig. 72).

The schist above the thrust plane is much more sericitic than that below the fault. The thrust surface is about 40 feet long and dips 40° to the southeast. At the south end, the thrust is truncated by a steep fault dipping 75° to the northwest (Pl. 4, fig. 2). Flow cleavage in the vicinity dips steeply to the southeast. Cleavage

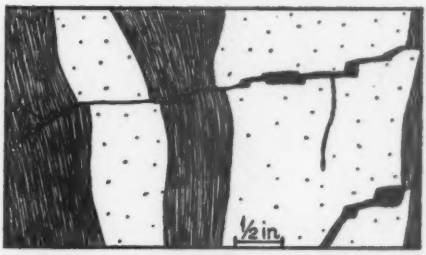



FIGURE 18. SMALL LATERAL DISPLACEMENT IN SANDY AND ARGILLACEOUS BEDS IN PETERS CREEK SCHIST

below the thrust plane is undisturbed, but it is dragged into a position of parallelism above the thrust.

The exposure suggests that the sericitic schist was thrust to the northwest over undisturbed schist. As stress continued and sufficient relief could no longer be attained laterally, the overriding schist buckled upward and finally fractured, thus producing the steeply dipping fault.

*Fault along fracture cleavage in Kell quarry.*—There is a small transverse fault along a fracture cleavage plane about 55 feet east of the west end of the slate exposure just above the northwest rim of Kell quarry. The fault strikes 02° and dips 70° east. It is a strike slip fault, the east block having moved northward a few inches with respect to the west block. The fault zone is several inches wide and contains much crushed slate and quartz vein material. The orientation of the fault with respect to slaty cleavage, and the directions of relative movement are indicated in Figure 19. The entire slate belt is offset similarly, but on a larger scale, about a third of a mile east of the Susquehanna River.

*Faulted anticline in Johnson quarry.*—A faulted northeast-plunging anticline was exposed in blasting operations near the north end of the southwest wall of Johnson slate quarry (operated by Funkhouser Company). The exposure (Pl. 4, fig. 4) was about 40 feet above the

floor of the quarry and 30 feet below the quarry rim. The slate within 12 feet of the fault was very much disturbed and injected by thick quartz veins.

The fault strikes about 70° and dips 81° to the north-northwest. It is marked by chloritic gouge a fourth to 2 inches thick. On the north, the fault is bounded by crushed and brecciated slate with numerous quartz stringers. The direction of movement is not easily determined, but this seems to be a reverse fault, the north side having moved up with respect to the south side. This longitudinal fault is only about 300 feet southeast of the major fault in the area.

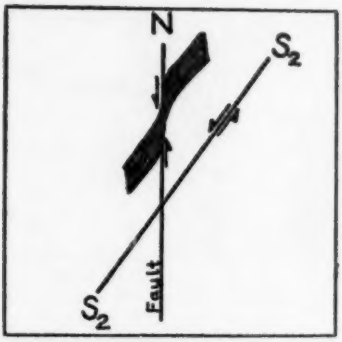



FIGURE 19. DIRECTION OF MOVEMENT ALONG SMALL LATERAL SHEAR FAULT AND $S_2$ IN KELL QUARRY

*Major fault in the area.*—The major fault in the Peach Bottom syncline cuts out the conglomerate and part of the slate for about 9 miles along the north side of the syncline. The fault zone is marked by deep weathering. There are no outcrops along its extent. On the east bank of the Susquehanna River, in the vicinity of the fault, there is a gap of about 60 feet between the north end of the exposed slate and the south end of the schist.

The displacement along the fault is estimated to be at least 450 feet (Behre, 1939, p. 367). The fault plane may dip southeast and parallel the axial plane of the overturned Peach Bottom syncline, or it may dip steeply to the northwest, as does the small parallel fault in Johnson quarry, a few hundred feet to the southeast (Pl. 4, fig. 4). The omission of beds along the fault suggests that the relative movement was down on the southeast side.

The fault may be genetically related to the horizontal or gently-dipping thrust planes. The major fault and the thrust surfaces may represent shear planes formed by the rotational force which overturned the syncline toward the northwest (Fig. 20). The gently-dipping shear planes formed earlier and underwent some degree of rotation before major faulting occurred. This view modifies somewhat a previous one expressed by the writer (Agron, 1950).

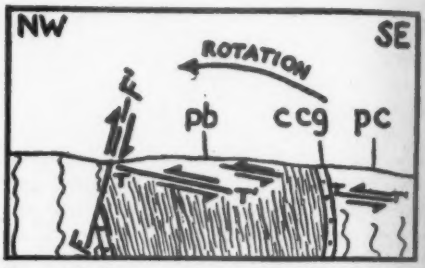



FIGURE 20. INTERPRETATION OF THE MAJOR FAULT ALONG THE NORTHWEST SIDE OF THE PEACH BOTTOM SYNCLINE

F-F′ fault, T-T′ thrust planes uniformly offset to northwest, pb-Peach Bottom slate, ccg-Cardiff conglomerate, pc-Peters Creek schist.

### *Structures in Kell Quarry*

Many structural features of deformed slate are well shown in Kell quarry (Fig. 21). The quarry is about 650 feet long, up to 160 feet wide, and about 45 feet deep. The quarry is not worked now. It can be entered on the northwest through a tunnel, which leads to Johnson quarry 200 feet to the southwest. Slaty cleavage strikes east-northeast and dips very steeply to the north. The quantity of good slate removed from the quarry was rather small because of the numerous crushed and badly disturbed zones which frequently contain irregular quartz veins. Fracture cleavage is strongly developed in a north-south band, about 150 feet wide, which lies near the western end of the quarry. The slate in the ledge above the northwest rim is badly crushed and injected by quartz veins. Several sets of lineations are present at that place (Pl. 5, fig. 4). A quartzitic bed or lens is exposed on a terrace south of the southeast rim of the quarry.

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| Figure | Structure | Formation | County | No. of readings | Contours (per cent) |
|---|---|---|---|---|---|
| A | Schistosity | Pb | Lancaster | 197 | 1-3-6-9-12-13 |
| B | Schistosity | Cegl | Lancaster | 30 | 5-10-20-30 |
| C | Schistosity | Pc | Lancaster | 282 | 1-4-8-12-16-20 |
| D | Schistosity | Pb | York | 228 | 1-5-10-15-26-22 |
| E | Schistosity | Cegl | York | 37 | 1-5-10-20 |
| F | Schistosity | Pc | York | 105 | 1-2-10-14-30-40-48 |
| G | Joints | Pb | Lancaster | 316 | 0.5-1-2-3-4-5-6 |
| H | Joints | Cegl | Lancaster | 99 | 1-2-4-6-8 |
| I | Joints | Pc | Lancaster | 504 | 0.5-1-2-3-4 |
| J | Joints | Pb | York | 276 | 1-2-3-4-5 |
| K | Joints | Cegl | York | 65 | 1-3-4-7 |
| L | Joints | Pc | York | 85 | 1-2-4-6-9 |
| M | Lin. and F.A. | Pb | Lancaster | 180 | 1-2-4-6-8-10-12 |
| N | Lin. and F.A. | Pc | Lancaster | 202 | 1-4-7-10-13 |
| O | Lin. and F.A. | Pb | York | 121 | 1-2-4-6-10-13 |
| P | Lin. and F.A. | Pc | York | 50 | 2-4-6-10-12-18 |
| Q | Fracture cl. | Pb | Lancaster | 131 | 0.5-1-3-5-7-9 |
| R | Fracture cl. | Pc | Lancaster | 82 | 1-2-4-6-8-9 |
| S | Fracture cl. | Pb | York | 58 | 1-4-7-12-17 |

T Synoptic diagram of schistosity, fracture cleavage and joints in Pb, Lancaster County.

U Synoptic diagram of schistosity, fracture cleavage and joints in Cegl, Lancaster County.

V Synoptic diagram of schistosity, fracture cleavage and joints in Pc, Lancaster County.

W Composite diagram of synoptic diagrams 2, 3, and 4.

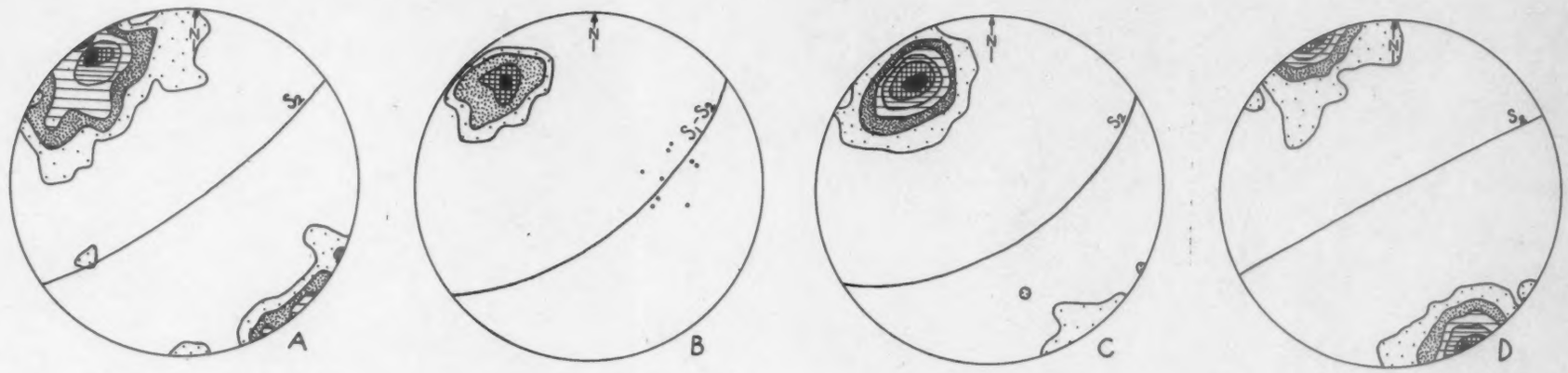
X Equal area net.

Pb—Peach Bottom slate.

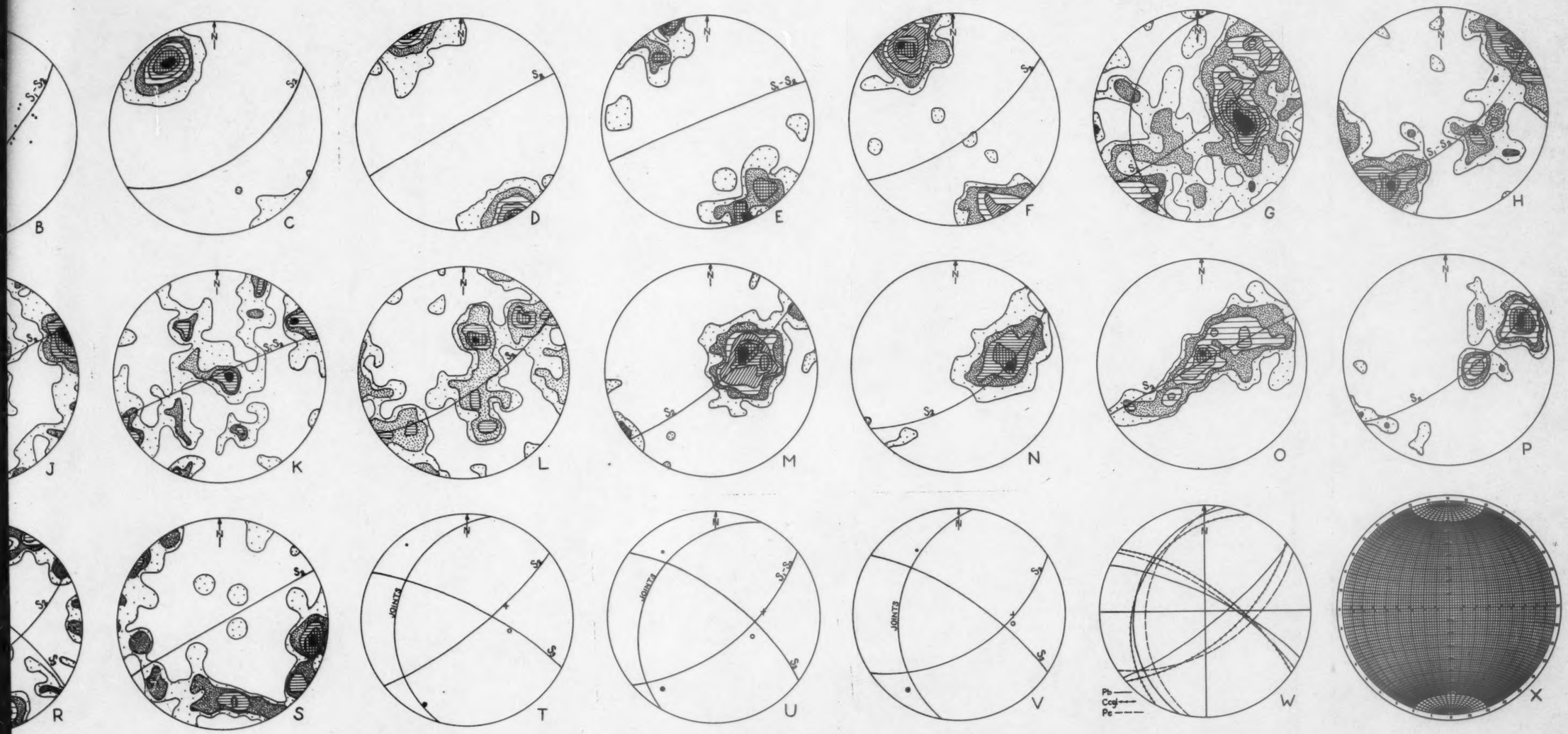
Cegl—Cardiff conglomerate.

Pc—Peters Creek schist.

York County diagrams also include data from Harford County.




STEREOGRAPHIC PROJECTIONS




**STEREOGRAPHIC PROJECTIONS OF FIELD STRUCTURES**

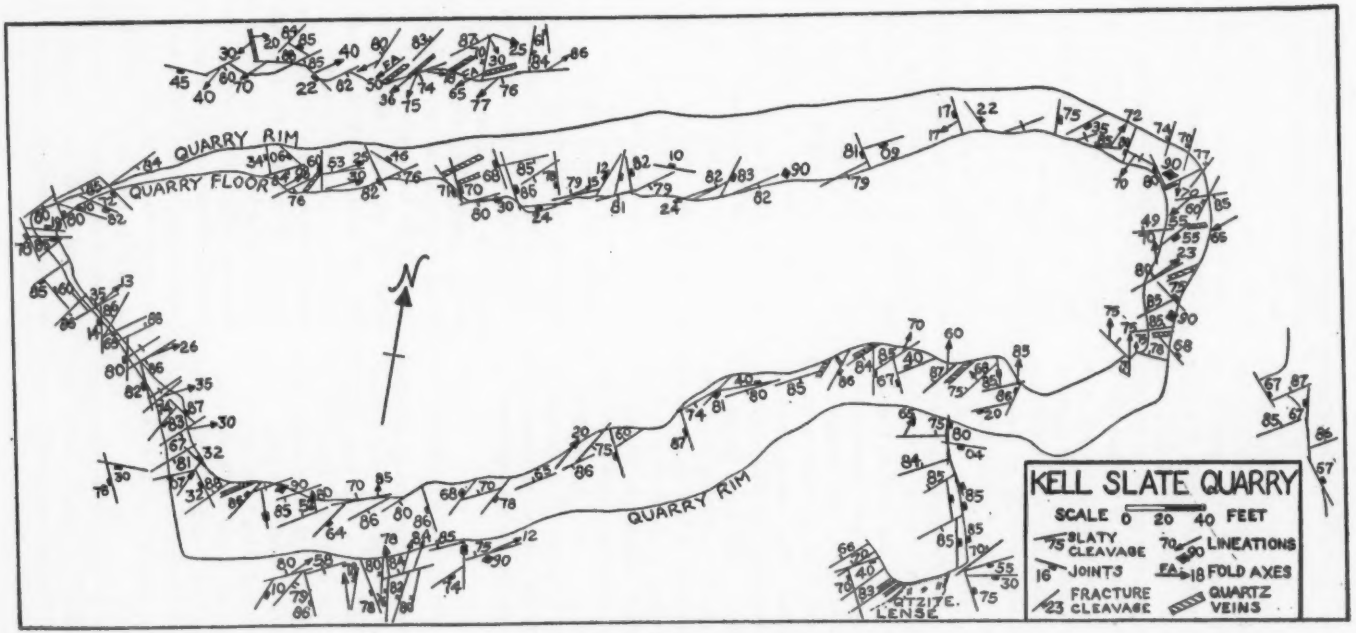



FIGURE 21.—KELL SLATE QUARRY

*Railroad Cut Along East Bank of Susquehanna River*

Inasmuch as the railroad cut along the east bank of the Susquehanna River affords the best section across the Peach Bottom syncline, the structures were mapped on a scale of 12½ feet to the inch in order that sufficient detail could be recorded. The map was reduced for presentation to a scale of 200 feet to the inch (Pl. 9). The base map of the railroad cut was taken from the electrification plans of the Columbia and Post Deposit Branch of the Pennsylvania Railroad, kindly made available to the writer by Mr. P. S. Settle of the Pennsylvania Railroad, Perryville, Maryland.

In order to present the abundant data on a scale of 200 feet to the inch, the various structures (joints, schistosity and bedding, lineations and fold axes, and fracture cleavage) were drawn individually along four sections, one alongside the other. The generally northeast trend of the schistosity (flow cleavage) is readily apparent. Lineations and fold axes show considerable variation in strike and plunge, but most of them plunge moderately to steeply northeast to southeast. A few plunge to the north, and still others to the southwest. Fracture cleavage, which roughly parallels the axial planes of folds, strikes more or less in the same direction as fold axes and lineations. Most of the joints are cross joints which lie approximately normal to the regional structures, and strike most frequently to the northwest or northeast.

## Drainage and Structure Pattern

Knopf and Jonas (1929) present an excellent discussion of the drainage and physiographic development of the region. The Susquehanna River is over a mile wide where it crosses the Peach Bottom syncline. It divides the area studied approximately in half. The major tributaries in the area are Fishing Creek and Peters Creek, in Lancaster County, Pennsylvania; Muddy Creek in York County, Pennsylvania; and Broad Creek in Harford County, Maryland. The streams are youthful and flow in deep, steep-walled valleys.

The writer tried to determine to what extent the seemingly dendritic drainage pattern may be controlled by schistosity and jointing in the underlying rocks. Compass rosettes were constructed from the strike readings for schistosity, jointing, and fracture cleavage. Readings east of the Susquehanna River were plotted separately from those west of the river because the drainage directions appear to be different in the two places.

Strike readings accompanied by dips of less than 45° (only a small percentage of all the readings) were not used because it was assumed that they are of relatively little importance in determining stream directions.

The drainage pattern of the area (Fig. 22) was traced from U. S. Geological Survey topographic sheets on a scale of 1: 62,500. An attempt was made to enclose the Peach Bottom syncline symmetrically, and to stay within the area where structural data had been collected.

The stream courses were broken down into their short component directions. The azimuth of each direction and its length on the map (in mm.) were measured. The lengths, converted to miles, were added up for 5-degree intervals of azimuth, and a drainage rosette was constructed. This could then be compared with the rosettes for the structural features (Fig. 23).

The results indicate that drainage is controlled by structures in the underlying rocks. The control is very good in York and Harford Counties, on the west side of the Susquehanna River, but appears to be rather poor in Lancaster County, on the east side of the river.

Poor corrélation in Lancaster County is perhaps to be expected, because there the Peach Bottom syncline and the slate ridge are not strong structural features, as they are west of the river. In Lancaster County, the syncline forms little more than a slender ribbon. In addition, structural readings in the eastern part of the area in Lancaster County are relatively scarce because the schist is largely sericitic and outcrops are very few. The structural statistics for the area in Lancaster County are too heavily weighted in favor of the sections near the Susquehanna River, where outcrops are abundant. Consequently, the available structural data are not representative for the drainage area in Lancaster County, as outlined in Figure 22.

Streams in the area west of the Susquehanna River flow in two main directions, northwest-southeast and northeast-southwest. The strong-

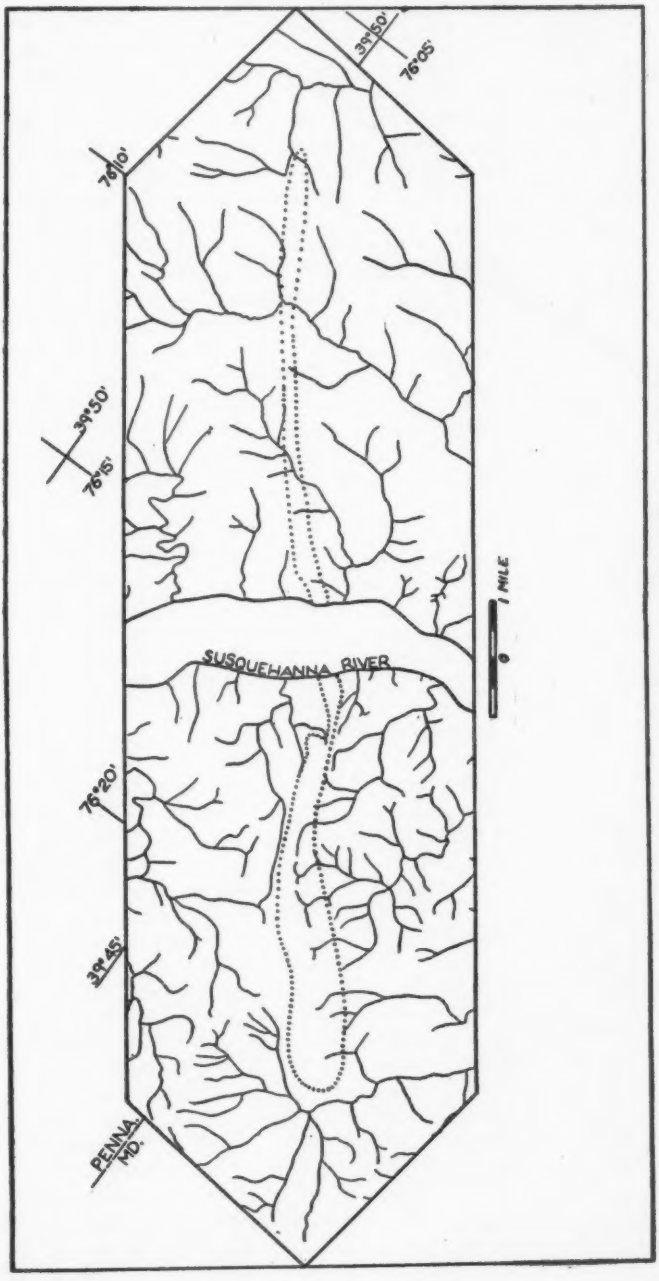



FIGURE 22.—DRAINAGE IN AREA OF PEACH BOTTOM SYNCLINE
Slate and conglomerate lie inside dotted line.....

er direction is to the southeast, towards the Susquehanna River. The north-east-southwest direction corresponds to the regional direction of schistosity; the northwest-southeast direction

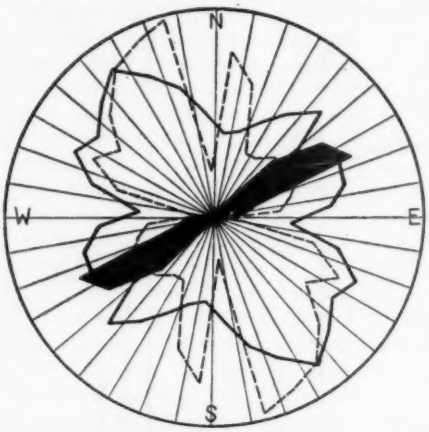



FIGURE 23. DRAINAGE AND STRUCTURES IN LANCASTER COUNTY

Black area—schistosity (flow cleavage); 1 in. = 112 readings.

Dashed line—joints and fracture cleavage; 1 in. = 28 readings.

Solid line—surface drainage; 1 in. = 8.4 miles of streams.

corresponds to the major direction of joints, which are the cross-joints of the area.

Although the schistosity is the most prominent structural feature in the rocks and represents the best parting plane, it exerts a smaller influence on stream direction than joints do. Peaks and lows in the rosettes of drainage and structural features are in fairly close agreement in the York County-Harford County area.

The total length of surface drainage in the 53.6 square-mile area measured in York and Harford Counties, as presented on the topographic maps, is 114 miles. This is equivalent to 2.2 miles of stream per square mile of area. In the 52 square-mile area measured in Lancaster County, the total drainage is 88 miles, or 1.7 miles of stream per square mile. This value is 74 per cent of that for the York and Harford County area.

## PETROFABRIC ANALYSIS

### *Introduction*

Forty-two petrofabric diagrams were prepared from ten oriented rock samples (Table 4). The diagrams are projections of *c*-axes of quartz and normals to mica cleavages in the lower hemisphere of an equal area net. Area percentages were counted at points a half centimeter or less apart. Most of the diagrams are oriented normal to *b* or B, and are viewed towards the east. An arrow and a small line indicate the strike and dip of the section. Angles of dip under 90° indicate a plane dipping towards the reader; angles greater than 90° indicate a plane dipping away from the reader.

One oriented section was prepared from sandy intercalations in the Peach Bottom slate, one from the Cardiff conglomerate, seven from the Peters Creek schist, and one from a quartz vein in the slate. There are eight muscovite diagrams, two chlorite, and one chloritoid diagram. A total of 2200 mica flakes were plotted, 400 in one diagram, 100 in two others, and 200 in each of the other mica diagrams.

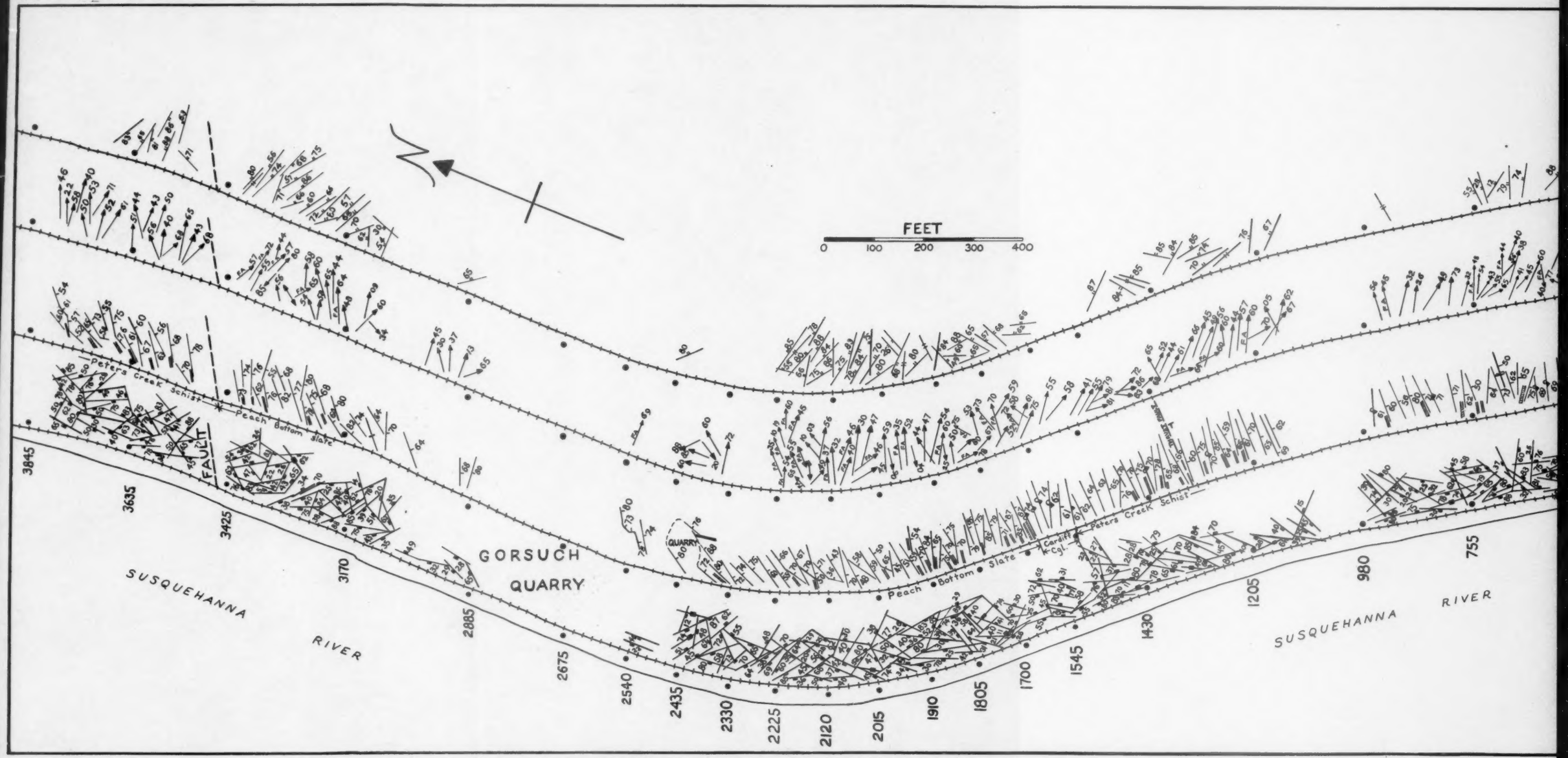
Thirty-one quartz diagrams were prepared, including selective diagrams (Knopf and Ingerson, 1938, p. 252) of large and small quartz grains, as well as collective diagrams. A total of 2400 *c*-axes of quartz were plotted. The number in each diagram varies from 400 to less than 100 (where the slide contained relatively few quartz grains).

### *Mica Orientation*

*General statement.*—The mica flakes are oriented in *ac* girdles of different degrees of completeness. More complete girdles have a weak maximum and one or more submaxima (Pl. 10, A2, A4), indicating several S-planes. Poorly developed girdles grade into a strong point maximum, which indicates the presence of a single S-surface ($S_2$).

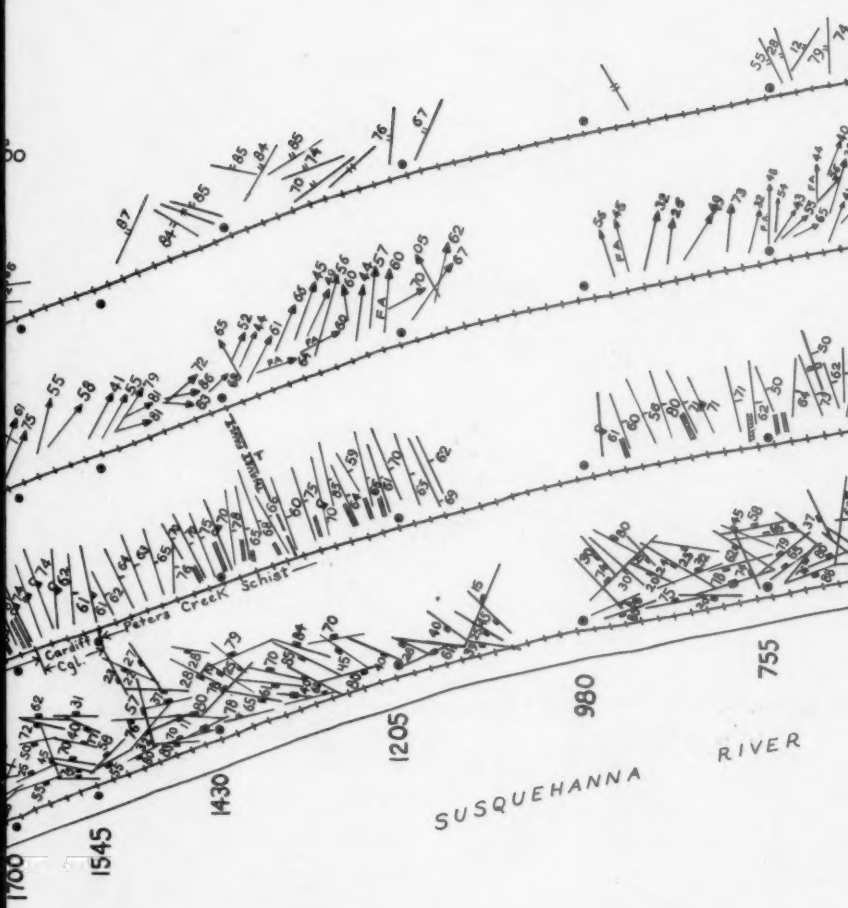
The peripheral extent of the 1 per cent and 4 per cent girdles and the height of maxima are presented in Table 5. In Figure 24, the height of maxima are plotted against the width of the 1 per cent and the 4 per cent girdles. Such data were plotted in a similar fashion by Cloos and Heitanen (1941, Figs. 22, 23).

The trend of the plotted points (Fig. 24) shows clearly that the strength of the maxima decreases as the peripheral extent of the girdle increases. The curve for the 1 per cent girdle shows a sharp change in slope near 270°. This agrees, in a general way, with the results obtained by Cloos in the Wissahickon schist, the



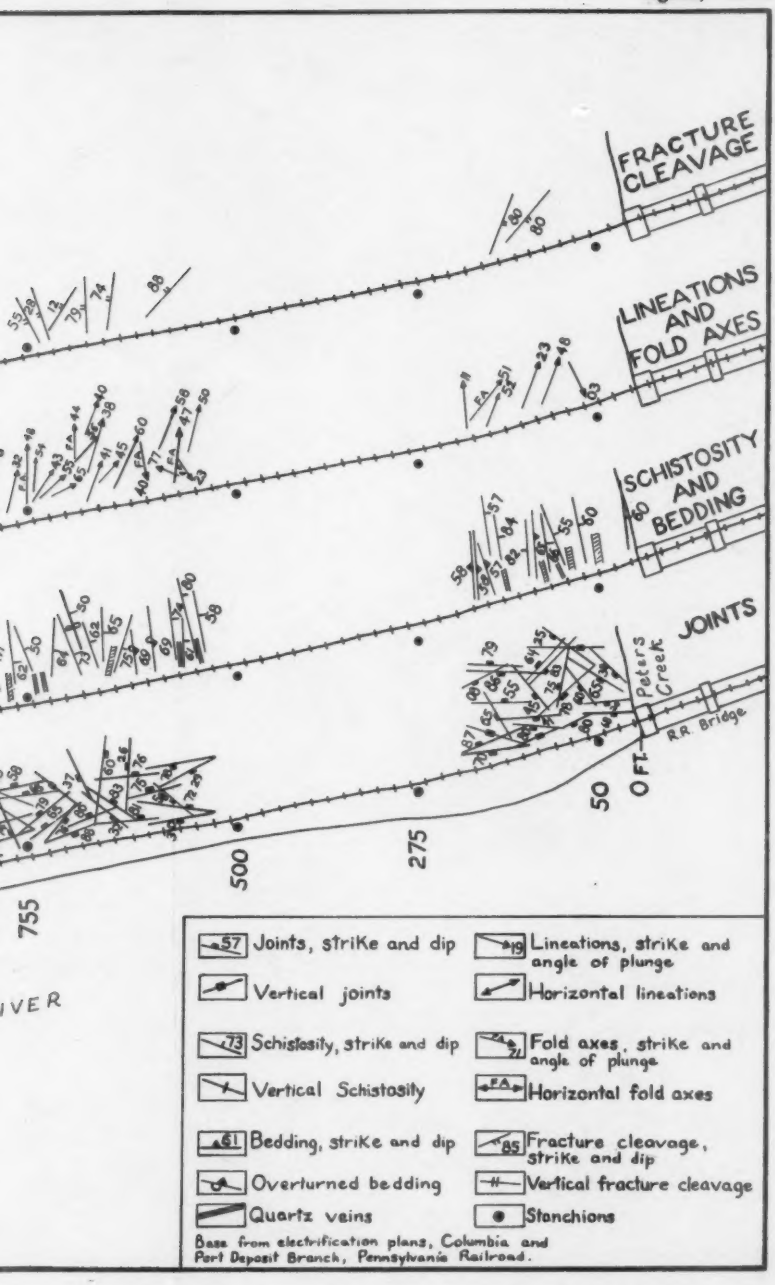

**STRUCTURAL FEATURES ALONG RAILROAD CUT NORTH OF PETERS CREEK**

The single track railroad follows the east shore of the Susquehanna River. The tracks are repeated above each other in order to show the several structures separately. Section crosses Peach Bottom syncline.




**CUT NORTH OF PETERS CREEK**

nna River. The tracks are repeated above
. Section crosses Peach Bottom syncline.


FRACTURE CLEAVAGE
LINEATIONS AND FOLD AXES
SCHISTOSITY AND BEDDING
JOINTS
Peters Creek
RR Bridge
0 FT
50
275
500
755
IVER
57 Joints, strike and dip
Vertical joints
73 Schistosity, strike and dip
Vertical Schistosity
61 Bedding, strike and dip
Overturned bedding
Quartz veins
19 Lineations, strike and angle of plunge
Horizontal lineations
71 Fold axes, strike and angle of plunge
FA Horizontal fold axes
85 Fracture cleavage, strike and dip
Vertical fracture cleavage
Stanchions
Base from electrification plans, Columbia and Port Deposit Branch, Pennsylvania Railroad.

Conestoga limestone, and the Antietam schist (1941, p. 41). He found a sharp break in the plots for the 1 per cent girdle at 210°. Perhaps this indicates that an imperceptible gradation between a point maximum arrangement of micas and a girdle orientation is not present, but rather the approximate division between the two can be recognized.

TABLE 4. ROCK SPECIMENS AND PETROFABRIC DIAGRAMS

| Spec. No. | Formation | Petrofabric diagrams | Location and remarks |
|---|---|---|---|
| 54 | Pb slate | A1-A2-A3-A4-A5 | Quartz layers in slate 3200 ft. NW. of RR. bridge over Peters Cr. |
| 111 | Ccgl | B1-B2-B3-B5 | Conglomerate exposure along E. shore of Susquehanna R. Good lineations. |
| 104 | Pc schist | C1-C2-C3-C5 | Quartzitic schist with micaceous partings. 1½ mi. SE. of Peach Bottom, York County, Penna. |
| 150 | Pc schist | D1-D2-D3-D5-D7 | Quartzitic and micaceous layers. Well foliated and lineated. 0.2 mi. SW. Peach Bottom, York County. |
| 164 | Pc schist | E1-E2-E3-E5 | Gneiss-like quartz-rich schist with thin micaceous layers. 30 ft. N. of RR. bridge over Peters Creek. |
| 165 | Pc schist | F1-F2-F3-F5-F7 | Arkosic quartz-rich schist at fork in road just east of auto bridge near mouth of Peters Creek. |
| 166 | Pc schist | G1-G2-G3-G5-G7 | Arkosic schist, forms cliff 775 feet NW. RR. bridge over Peters Cr. |
| 170 | Pc schist | H1-H2-H3-H4-H5 | Greenish schist in abandoned road metal quarry ½ mi. WNW. of McSparran, Lancaster County. |
| 174 | Pc schist | I1-I2-I3-I5 | Bottom of small tight syncline in quartzitic schist 560 feet NW. of RR. bridge over Peters Creek. |
| 52 | Qtz. vein in slate | J1 | 3-inch wide quartz vein in slate 1850 ft. NW. of RR. bridge over Peters Creek. |

*Interpretation of mica diagrams.*—[*A2 and A4,*] The most complete girdles, and hence the greatest degree of rotational deformation (Sander, 1930), were found in the slate specimen. The one per cent muscovite girdle (Pl. 10, diagram A2) occupies 326° along the periphery. The diagram shows a maximum of 13 percent and several submaxima, the largest of which is 11 per cent. The chloritoid diagram for the slate (A4) shows a 1 per cent girdle of 316°. It contains a maximum of 10.5 per cent, another of 10 per cent, and five submaxima ranging from 8.5 to 6 per cent.

The angular distance between the two strongest maxima is a little over 30° in each diagram. If the two muscovite maxima are rotated 15° counterclockwise (looking east), they are approximately superimposed on the two chloritoid maxima.

Slaty cleavage and bedding are both perpendicular to the 11 per cent muscovite maximum. The strongest chloritoid and muscovite maxima, however, lie about 30° from the normal to $S_1$–$S_2$, indicating that more flakes are inclined some 30° to $S_1$–$S_2$ than lie parallel to these planes. The plane corresponding to the chief maximum may be called $S_3$. As the $S_3$ and $S_3$ planes of muscovite do not coincide with the corresponding planes in the chloritoid diagram, the latter planes may be considered $S_3'$ and $S_3'$. Several smaller submaxima are distributed along the girdles in A2 and A4. The rather complete girdles indicate that the slate underwent much internal rotation during deformation.

*Diagram B2* shows the orientation of 200 chlorite flakes in the Cardiff conglomerate. They form a B-girdle of 286°, with a maximum concentration of 20 per cent. The hand specimen is poorly foliated, but has a distinct lineation. The schistosity plane ($S_2$) determined by the chlorite grains, is rotated 16° counter-clockwise from bedding as measured in the field. A minor concentration of cleavage poles lies near the top of the diagram, about 83° from the

pole of $S_2$. This small concentration is at the pole of the BC shear joints ($S_3$), which intersect bedding in a lineation. This 2 per cent concentration indicates that at least a few chlorite flakes were oriented along the BC joints, either by shear or by later mimetic crystallization. The lineations are due to the intersection of $S_1$ with $S_2$, and also $S_3$ with $S_2$.

peripheral spread of only 160°. An S-tectonite with one predominating shear plane is indicated. The quartz diagrams (E1, E3, E5), however, suggest rotation about $a$ (= B′). Although the quartz indicates that the rock became an R-tectonite at a later stage, the micas were apparently not affected.

*F2.* The muscovite in F2 shows a tendency to

TABLE 5. WIDTH OF MICA GIRDLES AND HEIGHT OF MAXIMA

| Diagram | Mineral | No. of grains | 1 per cent girdle (degrees) | 4 per cent girdle (degrees) | Height of maxima (per cent) |
|---|---|---|---|---|---|
| A2 | Muscovite | 100 | 326 | 164 | 13–11–7–6 |
| A4 | Chloritoid | 200 | 316 | 146 | 10.5–10–8.5–8–6.5 |
| B2 | Chlorite | 200 | 286 | 148 | 20.5–2 |
| C2 | Muscovite | 200 | 307 | 180 | 15–9.5–7.5–5.5 |
| D2 | Muscovite | 200 | 264 | 106 | 22–3.5–2.5 |
| E2 | Muscovite | 200 | 160 | 104 | 20 |
| F2 | Muscovite | 400 | 214 | 126 | 16–12 |
| G2 | Muscovite | 200 | 163 | 107 | 20 |
| H2 | Muscovite | 200 | 208 | 133 | 15–14 |
| H4 | Chlorite ..... | 100 | 255 | 129 | 16–3 |
| I2 | Muscovite | 200 | 301 | 156 | 13–7.5–7–6–2.5 |

*Diagram C2* shows 200 muscovites in a quartz-sericite bed in the Peters Creek schist. The muscovites form a good girdle of 307°. The 15 per cent maximum is the pole of megascopic schistosity. A 9.5 per cent maximum is the pole of $S_2'$, which is 26° counter-clockwise from $S_2$. Another plane ($S_2''$) to which some of the quartz orientations seem to be related, can be distinguished about 20° beyond $S_2'$. The succession of S-planes have been rotated in a counter-clockwise direction, looking east.

*D2* shows a muscovite girdle around B′, which plunges 77° to the east, B′ has been rotated about $c$ to within 13° of the $a$ axis. Most of the muscovites lie in the $S_2$ plane, as indicated by the 22 per cent maximum. Some muscovite flakes have been dragged into a fracture cleavage plane ($S_3$) which lies 40–50° counter-clockwise from $S_2$. A smaller number of muscovite flakes follow a second fracture cleavage plane ($S_4$), which lies about 77° on the other side of $S_2$.

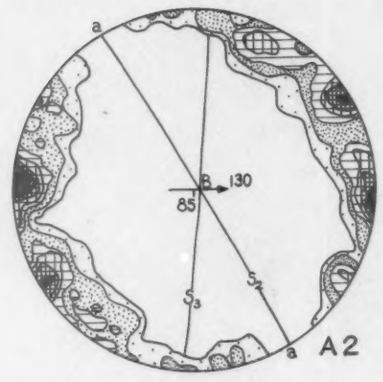
The muscovites lying in $S_3$ were sheared into that plane without breaking, but flakes which were oriented into $S_4$ were bent and fractured. Later mimetic crystals of muscovite grew along $S_4$. The micas lying in $S_2$ are arranged en echelon due to slip along $S_3$. The foliation measured in the hand specimen (illustrated in the sketch in $D_2$) is the mean direction between $S_2$ and $S_3$.

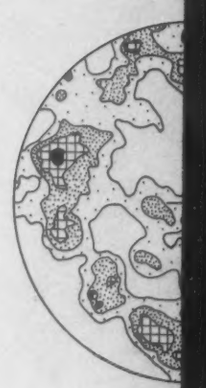
*E2* is a section normal to $a$. The micas show a distinct point maximum of 20 per cent and have a form a girdle around B. A 12 per cent submaximum lies 18° counter-clockwise from the 16 per cent maximum. The rock, a quartz-albite-muscovite schist, shows a poor foliation which lies midway between $S_2$ and $S_2'$. A third plane ($S_3$) is suggested by the four per cent submaximum which lies 48° counter-clockwise from $S_2$.

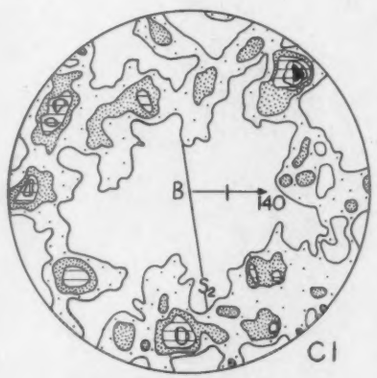
*G2* shows a strong 20 per cent maximum which is the pole of $S_2$. The micas extend along the periphery for only 163°, an arc hardly large enough to be considered an incipient girdle. However, the 4 per cent submaximum located 35° from the 20 per cent maximum, suggests that another S-plane ($S_1$) is present. Visible evidence of this is furnished by albite grains which are fractured along directions subparallel to $S_2$. The relative displacement of the several parts of the fractured albites indicates clockwise rotation about B. The quartz diagrams further suggest that this is a B $\wedge$ B′-tectonite, but the mica diagram gives no indication of it.

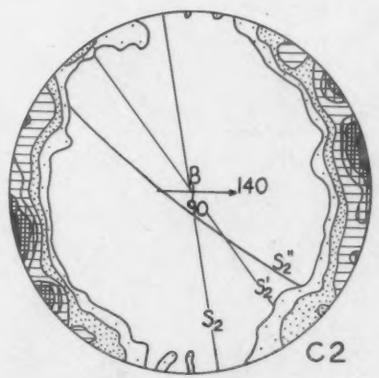
*H2 and H4* are muscovite and chlorite diagrams of epidotized Peters Creek schist which contains equal amounts of quartz, albite, and micas. Chlorites show a more complete girdle (255°) than muscovites (208°), which suggests chlorite is older and has undergone more rotation. The chlorite maximum lies between the 14 and 15 per cent muscovite maxima, which are due to slight waviness and crumpling of $S_2$. Megascopic schistosity lies normal to the 16 per cent chlorite maximum and midway

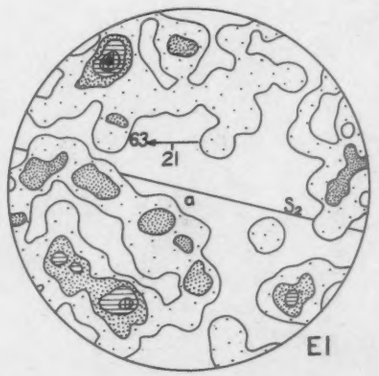
| Diagram | Mineral | No. of grains | Contours (per cent) |
|---|---|---|---|
| A1 | Quartz, large | 100 | 1-2-3-4-5 |
| A2 | Muscovite | 100 | 1-2-4-6-8-11-13 |
| A3 | Quartz, small | 100 | 1-3-4-5-8 |
| A4 | Chloritoid | 200 | 1-2-4-6-8-10 |
| A5 | Quartz, collective | 200 | 1-2-3-4 |
| B1 | Quartz, large | 100 | 1-3-5-7 |
| B2 | Chlorite | 200 | 1-2-4-8-15-20 |
| B3 | Quartz, small | 100 | 1-3-4-5-6 |
| B5 | Quartz, collective | 200 | 1-2-3-4-5 |
| C1 | Quartz, large | 200 | 1-2-3-4-5 |
| C2 | Muscovite | 200 | 1-2-4-7-9-14-15 |
| C3 | Quartz, small | 200 | 1-2-3-4-5 |
| C5 | Quartz, collective | 400 | 1-2-2.5-3-4 |
| D1 | Quartz, large | 100 | 1-3-4-5 |
| D2 | Muscovite | 200 | 1-2-3-8-15-20-22 |
| D3 | Quartz, small | 100 | 1-3-4-5-7 |
| D5 | Quartz, collective | 300 | 1-2-3-3.5 |
| D7 | Quartz, elongated grains | 100 | 1-3-5-6-7 |
| E1 | Quartz, large | 100 | 1-3-5-6-8 |
| E2 | Muscovite | 200 | 1-2-4-7-10-15-20 |
| E3 | Quartz, small | 100 | 1-3-5-6-8 |
| E5 | Quartz, collective | 200 | 1-2-3-4-5 |

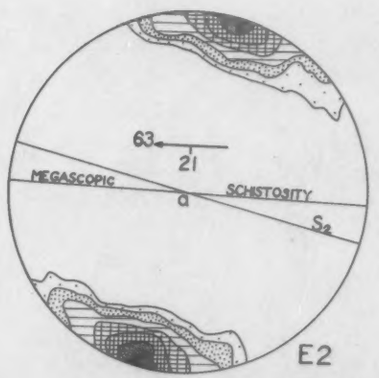






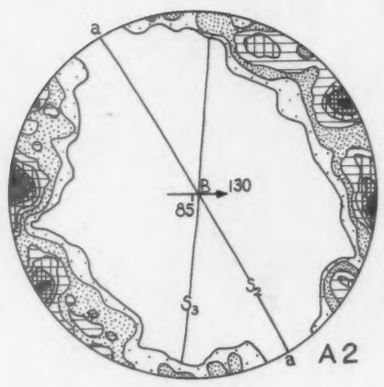



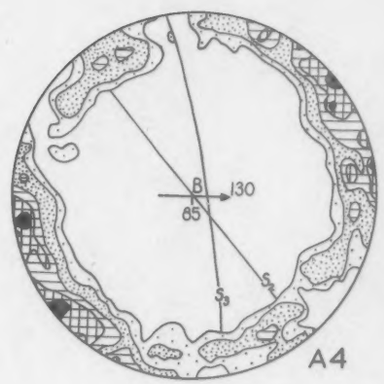


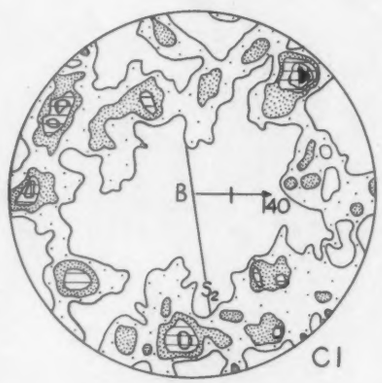



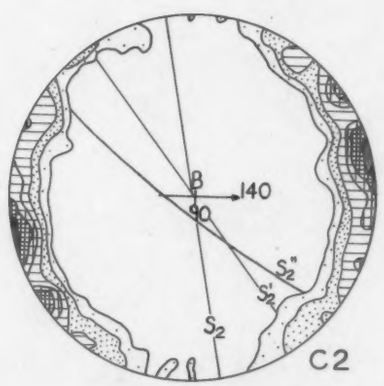
PETROFABRIC DIAGRAMS

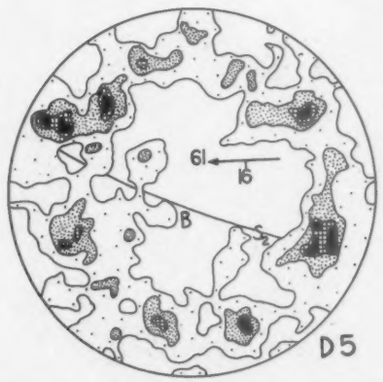
| Diagram | Mineral | No. of grains | Contours (per cent) |
|---|---|---|---|
| A1 | Quartz, large | 100 | 1-2-3-4-5 |
| A2 | Muscovite | 100 | 1-2-4-6-8-11-13 |
| A3 | Quartz, small | 100 | 1-3-4-5-8 |
| A4 | Chloritoid | 200 | 1-2-4-6-8-10 |
| A5 | Quartz, collective | 200 | 1-2-3-4 |
| B1 | Quartz, large | 100 | 1-3-5-7 |
| B2 | Chlorite | 200 | 1-2-4-8-15-20 |
| B3 | Quartz, small | 100 | 1-3-4-5-6 |
| B5 | Quartz, collective | 200 | 1-2-3-4-5 |
| C1 | Quartz, large | 200 | 1-2-3-4-5 |
| C2 | Muscovite | 200 | 1-2-4-7-9-14-15 |
| C3 | Quartz, small | 200 | 1-2-3-4-5 |
| C5 | Quartz, collective | 400 | 1-2-2.5-3-4 |
| D1 | Quartz, large | 100 | 1-3-4-5 |
| D2 | Muscovite | 200 | 1-2-3-8-15-20-22 |
| D3 | Quartz, small | 100 | 1-3-4-5-7 |
| D5 | Quartz, collective | 300 | 1-2-3-3.5 |
| D7 | Quartz, elongated grains | 100 | 1-3-5-6-7 |
| E1 | Quartz, large | 100 | 1-3-5-6-8 |
| E2 | Muscovite | 200 | 1-2-4-7-10-15-20 |
| E3 | Quartz, small | 100 | 1-3-5-6-8 |
| E5 | Quartz, collective | 200 | 1-2-3-4-5 |

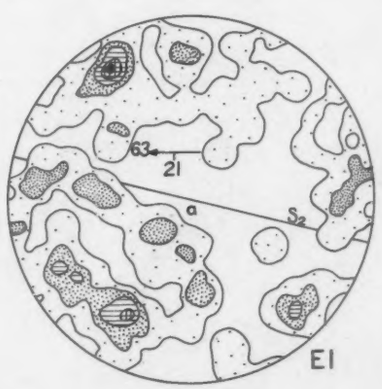


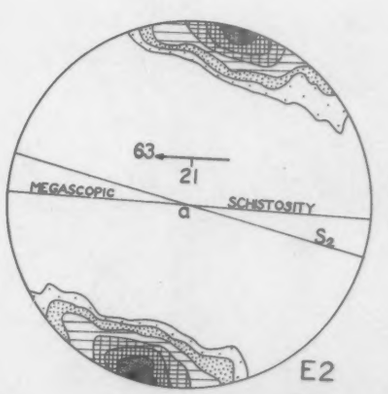


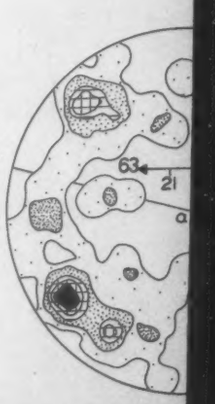


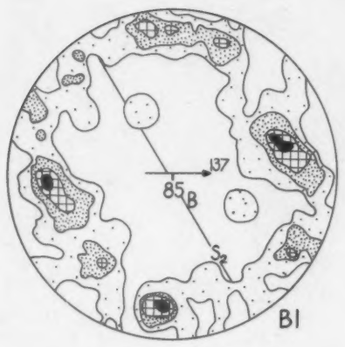


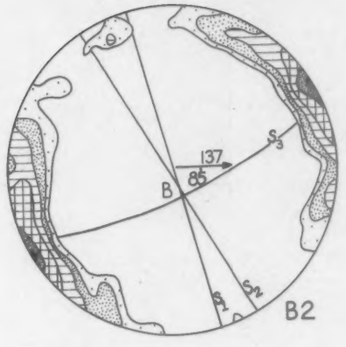


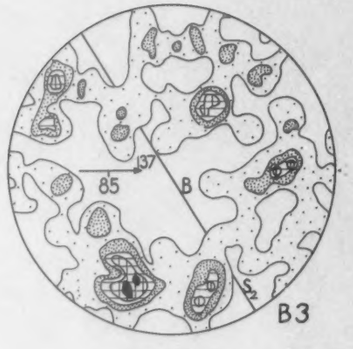



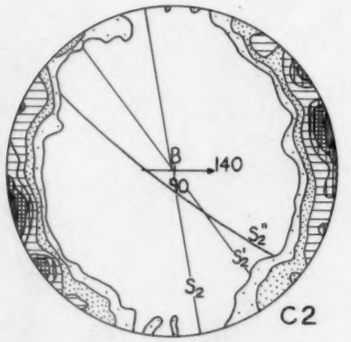



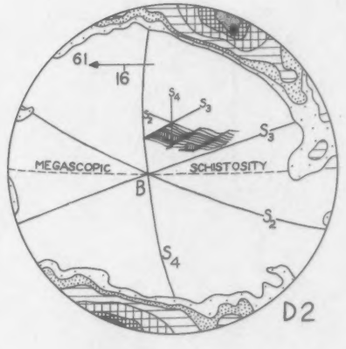

PETROFABRIC DIAGRAMS

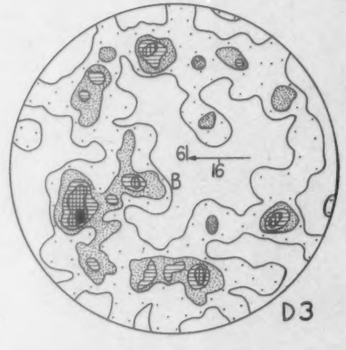


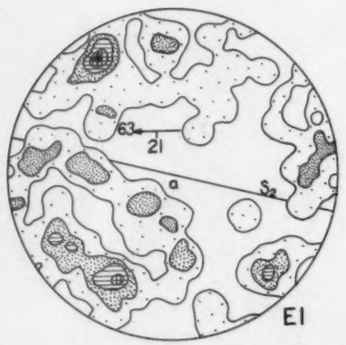



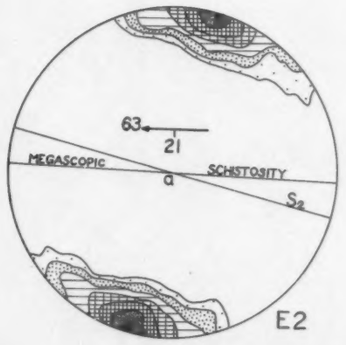



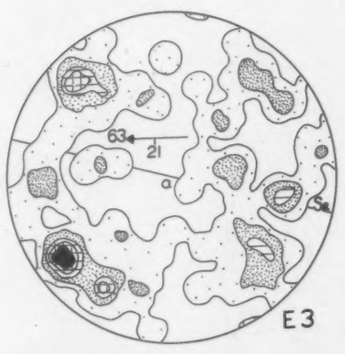


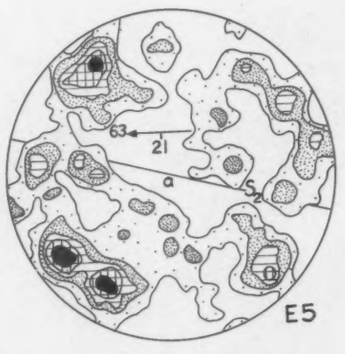





**PETROFABRIC DIAGRAMS**

FIGURE 24.—WIDTH OF MICA GIRDLES AND HEIGHT OF MAXIMA

between $S_2$ and $S_2'$ of muscovite, as might be expected. A fracture cleavage ($S_4$) is followed by very few micas but it is reflected in the quartz diagrams.

*I2* was prepared from quartzitic Peters Creek schist at the center of a small tightly folded syncline. The diagram shows a fairly complete muscovite girdle of 301°. The girdle, with its several weak

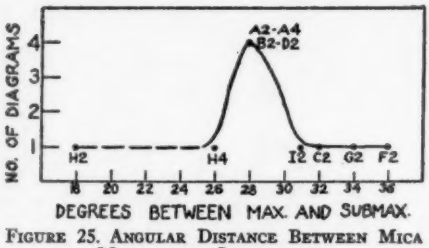



FIGURE 25. ANGULAR DISTANCE BETWEEN MICA MAXIMA AND SUBMAXIMA

submaxima, is formed by muscovites which lie parallel to the folded S-plane. Folding has distributed the muscovite poles along the periphery. Thus, it becomes possible for an h01 fracture cleavage plane to build up the strongest maximum (13 per cent) although only a small percentage of the total muscovites lie parallel to this plane.

The rotational deformation indicated by mica girdles is continuous, but it does not proceed at a uniform rate or intensity. The maximum and submaximum in a mica diagram indicate the strongest S-planes of the many which may have rotated through the rock. The writer measured the angular distance between the maximum and submaximum in diagrams showing more than a single point maximum. A distribution curve (Fig. 25) was constructed, in which the angular distance between maximum and first submaximum is plotted against the number of diagrams in which that distance occurs.

The angular distance between the maximum and submaximum ranges from 26° to 36° in all but one of the mica diagrams, and in that one it is 18°. The curve (Fig. 25) rises to a sharp peak at 28°, and there is a suggestion of a fairly symmetrical distribution about this peak. While the distance between maximum and submaximum is 28° in four of the diagrams, it is a different value in each of the other diagrams. The arithmetic average for the distance, in all diagrams, is 28.9°. Although the number of diagrams studied is limited, there is nevertheless an indication that rotational deformation proceeded through many degrees of arc. However, Figure 25 suggests that the deformation reached two peaks of intensity approximately 29° apart.

*Regional arrangement of mica poles.*—Diagram K2 shows the mica maxima (dots) and submaxima (circles) plotted in the geographic plane. They fall more or less along a plane perpendicular to the regional trend of schistosity ($S_2$). That is, the cleavage poles of micas lie along the trace of the *ac* plane. The larger concentration of mica poles in the northwest quadrant is due to the fact that the regional schistosity dips predominantly to the southeast. The distribution of points elsewhere along the great circle (*ac* plane) is due to flexures and folding of schistosity planes, as well as internal rotation of S-planes.

Although some of the quartz diagrams show rotation about B′ as well as B, the axis of rotation in the mica diagrams is almost always B. The micas were affected most by the earlier and stronger rotation, which was about B. Quartz, being a much more sensitive indicator of deformation than mica, shows the effects of later, less intense stages of deformation which are not so well indicated by the micas.

### *Quartz Fabric*

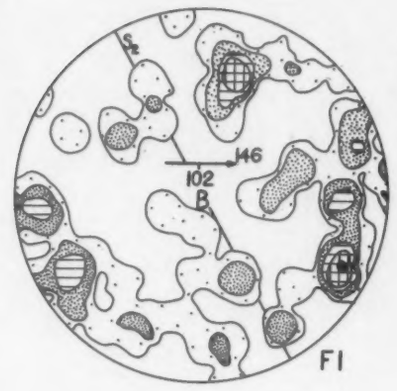
*General statement.*—Quartz diagrams are much more difficult to interpret than mica diagrams, which portray rotation of S-planes or slippage along the single cleavage plane of mica. Quartz responds to stress much more readily than mica, but its orientation mechanism is not so well understood. According to Sander (1930), and Griggs and Bell (1938), orientation in quartz results from fracturing along definite crystallographic directions, followed by alignment of the splinters in the directions of movement. Hietanen (1938, p. 38) and other Finnish geologists believe that quartz orientation may be due to gliding along (0001) and rupture along (0001) when external rotation of the grain cannot take place. Other theories have also been proposed, but "... there is as yet no accepted theory of a general nature which will expalin all the known fact of its (quartz's) orientation." (Fairbairn, 1942, p. 74).

The writer has prepared 21 partial diagrams of large and small grains, or grains selected for other reasons, and 10 collective diagrams, each

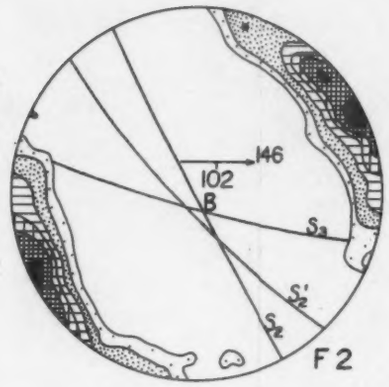
V6

1/8/0

| Diagram | Mineral | No. of grains | Contours (per cent) |
|---|---|---|---|
| F1 | Quartz, large | 50 | 1-3-5-7-10 |
| F2 | Muscovite | 400 | 1-2-4-6-8-12-16 |
| F3 | Quartz, small | 140 | 1-2-3-4-5 |
| F5 | Quartz, collective | 200 | 1-2-3-4-5 |
| F7 | 23 c-axes in one shattered quartz, grain | | 4-8-16 |
| G1 | Quartz, large | 70 | 1-3-5-7-8 |
| G2 | Muscovite | 200 | 1-3-4-10-15-20 |
| G3 | Quartz, small | 103 | 1-3-5-6-7 |
| G5 | Quartz, collective | 276 | 1-2-3-4-5 |
| G7 | Quartz in syntectonic vein | 103 | 1-3-5-7-9 |
| H1 | Quartz, large | 100 | 1-3-5-7 |
| H2 | Muscovite | 200 | 1-2-4-7-10-14-15 |
| H3 | Quartz, small | 100 | 1-3-5-7 |
| H4 | Chlorite | 100 | 1-3-5-8-10-12-16 |
| H5 | Quartz, collective | 200 | 1-2-3-4-5 |
| I1 | Quartz, large | 100 | 1-3-5-7-8 |
| I2 | Muscovite | 200 | 1-2-4-6-9-12-13 |
| I3 | Quartz, small | 100 | 1-3-5-7 |
| I5 | Quartz, collective | 200 | 1-2-3-4-5 |
| J1 | Needle quartz in vein | 200 | 1-5-10-15-18 |
| K1 | Synoptic diagram of quartz maxima in 39 diagrams | | 1-2-3-4-4.5 |
| K2 | Synoptic diagram of mica maxima in 11 diagrams, horizontal projection. | | |
| K3 | Synoptic diagram of quartz maxima in D1, D3, D5, D7. | | |
| K5 | Synoptic diagram of quartz maxima in A1, A3, A5. | | |



F1

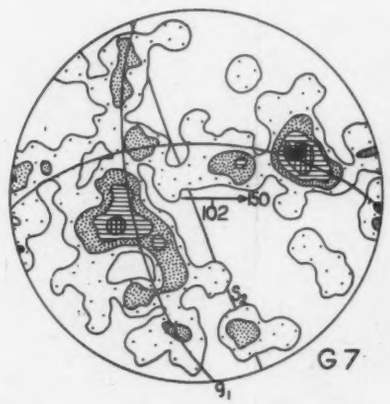


F2

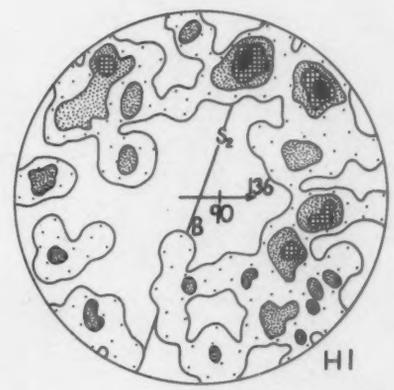
F3

F5

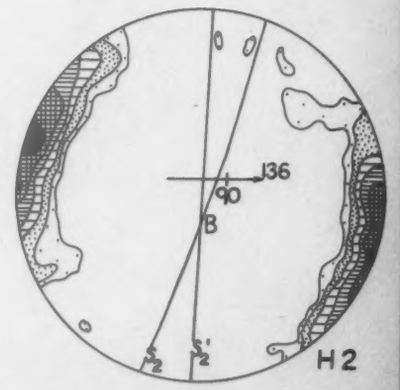
G5



G7



H1



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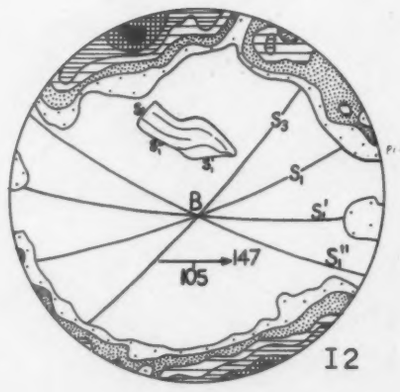

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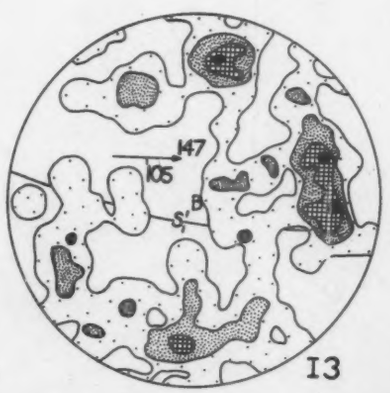

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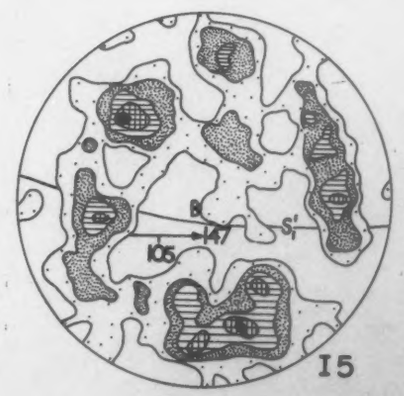

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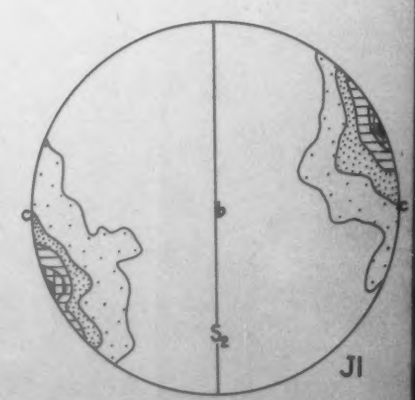
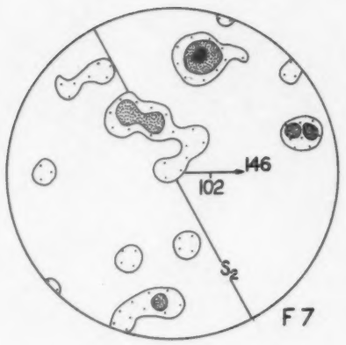


J1

F2

F3

F5



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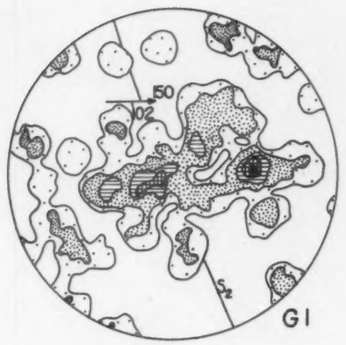

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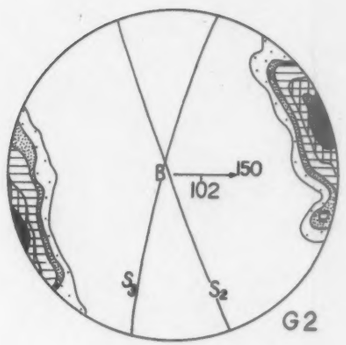

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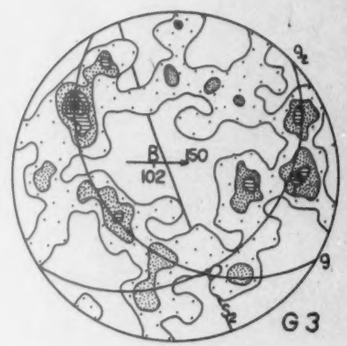
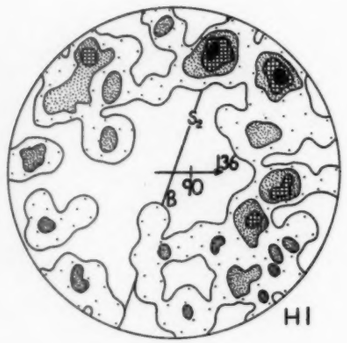


G3

G7



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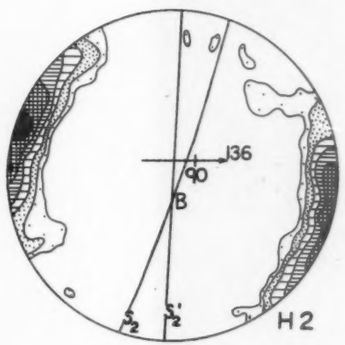

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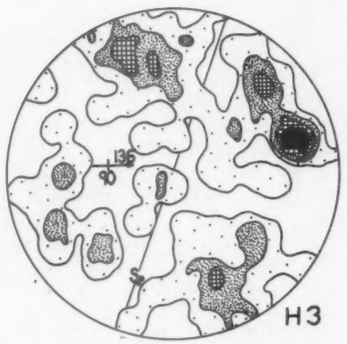

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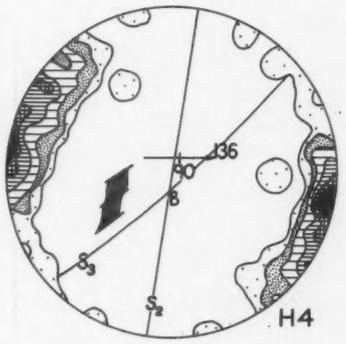

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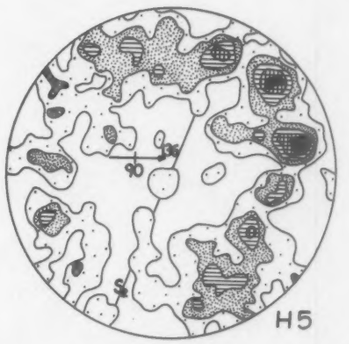

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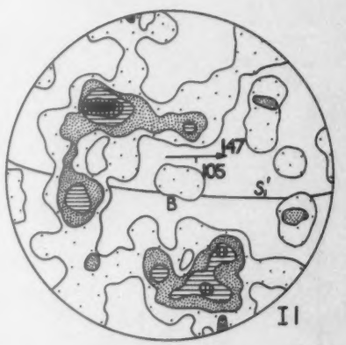

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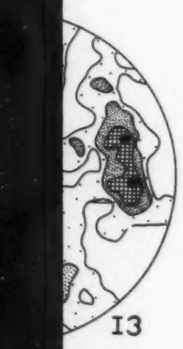

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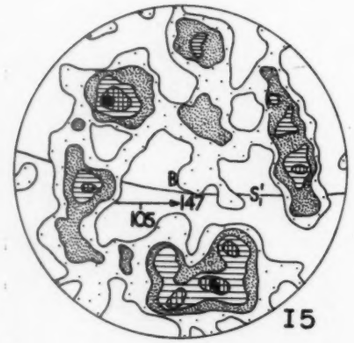

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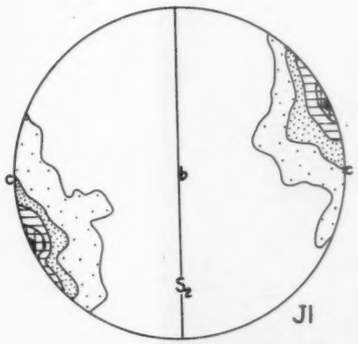
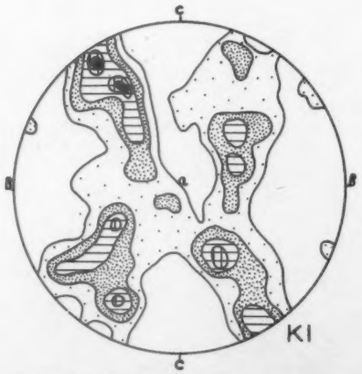
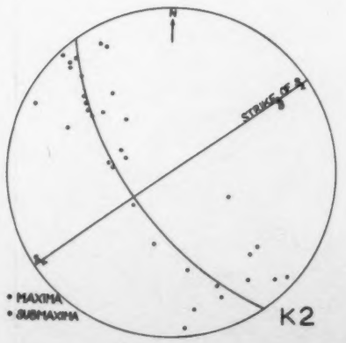


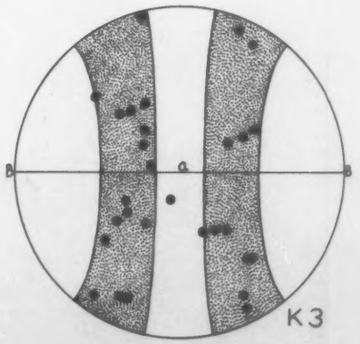
J1



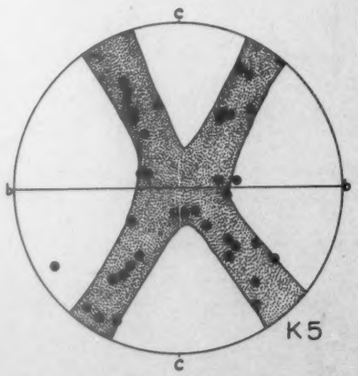
K1



K2



K3



K5

**PETROFABRIC DIAGRAMS**

repr
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representing all the quartz grains in a thin section. Large quartz grains are much strained and do not show the high degree of recrystallization displayed by small grains. They are thus less completely adjusted than small quartz grains. The partial diagrams for small grains may be expected to show the deformational effects of a stage somewhat later than that represented in the large-quartz diagrams.

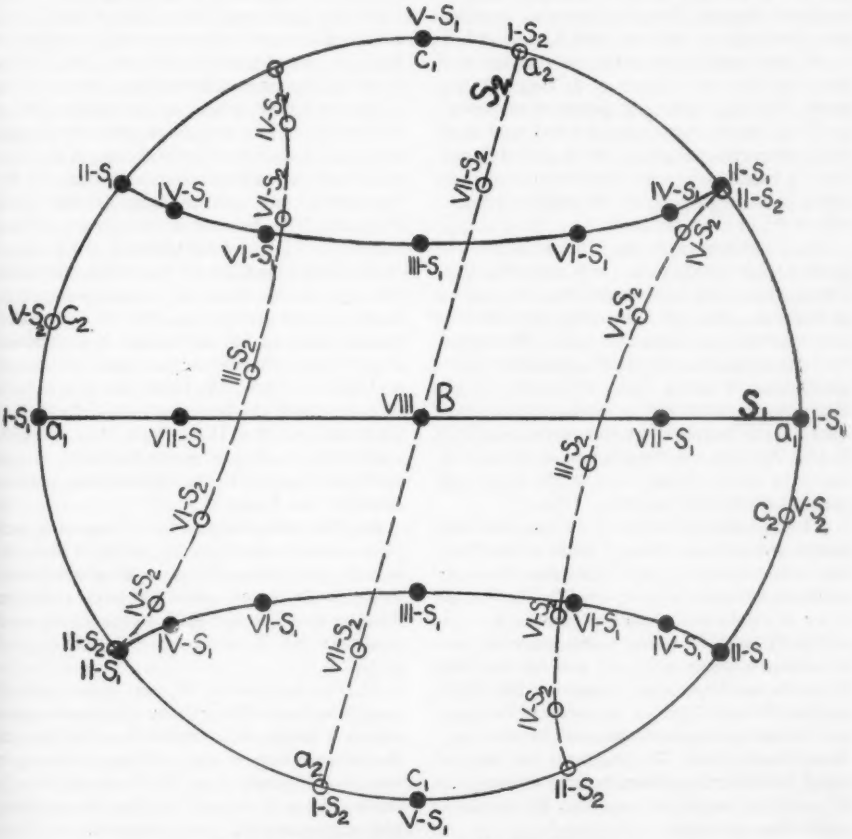


FIGURE 26.—DESIGNATION OF QUARTZ MAXIMA
Modified after Hietanen (Cloos and Hietanen, 1941).

Positions of quartz maxima (I to VIII) are designated according to Griggs and Bell (1938, Fig. 8) and Fairbairn (1942, Fig. 123). In addition, the S-plane to which the maximum is related is sometimes appended to the number of the maximum, as is done by Hietanen (1941, p. 168). Thus, V-$S_3$ is a V maximum lying normal to $S_3$, VI-$S_3$ is the VI position with respect to an *ab* plane, $S_3$, and so on (Fig. 26).

Most of the quartz diagrams show an irregular *ac* girdle which is not quite peripheral, but lies on a small circle about 20° in from the periphery. Fairbairn (1939, p. 1487) has found this type of girdle to predominate in certain Quebec quartzites.

*Interpretation of Quartz Diagrams. A1* (Pl. 10) a projection of large quartz grains in sandy slate, shows a fairly complete *ac* girdle which appears to have a monoclinic symmetry, with Bc as the symmetry plane. The 5 per cent maximum and the 4 per cent submaxima lie in the IV and VI positions

corresponding to $S_3$, as located in the mica diagram (A2). The large maximum in the southwest quadrant lies very near the VII-$S_3$ position, and so does one of the smaller 3 per cent submaxima in the northeast quadrant. Concentrations other than these two, show better correlation with $S_2$ than with $S_3$.

*A3.* The small quartz grains are oriented much more strongly with respect to $S_3$ than the large grains. The larger 8 per cent maximum corresponds to IV-$S_3$, but the other maximum and most of the large submaxima still occupy IV-$S_2$ and VI-$S_2$ positions. A triclinic symmetry has been imposed on the small grains in place of the monoclinic symmetry still shown by the large grains.

The 8 per cent maximum also lies in the VIII position with respect to B′ (= *a*, approximately). The maximum lies between the periphery and the B′ lineations. Thus, the *c*-axes which form the 8 per cent maximum are subparallel to B′. This orientation may be due to dimensional rolling of the quartz grains about B′ during plastic flow of the enclosing slate. Sander attributes a similar orientation of quartz grains in calcite to such a mechanism (1930, D 176). Hietanen ascribes this type of quartz orientation to gliding along a face in the prism zone (parallel to *ab*) (1938, p. 109).

*A5.* The collective diagram of 200 large and small quartz grains clearly shows a girdle around B. It also indicates, perhaps even better than the partial diagrams, a *bc* girdle around *a* (= B′). The B′ girdle is not so well formed as the girdle around B.

The IV and VI maxima positions are the most prominent in diagrams A1, A3, and A5. According to the fracture hypothesis of Griggs and Bell (1938), maxima IV and VI are due to rupture of the quartz into needlelike fragments bounded by the r or z rhombohedral faces. The fragments are then oriented with their long direction ([r:z] for maximum IV, and [r:r] or [z:z] for maximum VI) parallel to *a*, the direction of slip.

*B1.* The large quartz grains in the Cardiff conglomerate form a good B girdle which, however, is somewhat elongated in the *a* direction. A weak monoclinic symmetry is present. The locations of the three 7 per cent maxima are governed by $S_2$ and $S_3$. The 7 per cent maximum near the bottom of the diagram lies at VI-$S_3$ and also between positions IV-$S_2$ and VI-$S_2$. The 5 per cent submaximum a short distance to the northwest lies about 12° from VII-$S_3$. The elongated 7 per cent maximum at the west end of the diagram lies about 12° from IV-$S_3$ and about 15°, counter-clockwise, from a line joining IV-$S_2$ and VI-$S_2$. The other elongated 7 per cent maximum, at the east side of the diagram, lies at VII-$S_3$. The small 5 per cent submaximum below it lies between VI-$S_3$ and IV-$S_3$, and also near IV-$S_2$. The string of 5 per cent submaxima in the northeast quadrant passes through II-$S_2$, and also approaches V-$S_3$ at the left and IV-$S_2$ at the right. The maxima are dragged out along the girdle.

*B3.* The small quartz grains show a B girdle, but it is less distinct and not so close to the periphery as the girdle formed by the large quartz (B1). The girdle in B3 also shows a lower degree of symmetry. Many axes appear to have moved towards the center from the upper part of the diagram. The maxima are all either double or dumbbell-shaped, as if they had been stretched due to rotation around *a* (= B′). The double 6 per cent maximum lies near VI-$S_3$, VI-$S_2$, and IV-$S_3$. It extends upward towards III-$S_2$. The double 5 per cent submaximum at the bottom of the diagram lies about 12° from III-$S_3$ and VII-$S_2$. The other double 5 per cent submaximum, in the eastern part of the diagram, lies 10° from VII-$S_3$. Slightly above it and to the left, is a dumbbell-shaped 5 per cent submaximum which falls exactly on VI-$S_2$ and VI-$S_3$. The lower part of the double submaximum in the northwest part of the diagram lies in the vicinity of IV-$S_2$, IV-$S_3$, II-$S_2$, and II-$S_3$. Concentrations of *c*-axes are much stronger in areas which are common to standard maxima positions related to two S-planes.

*B5.* The collective diagram of large and small quartz grains is similar to B1, in that it presents a B girdle elongated parallel to *a*. B5 also suggests a weaker girdle, inclined about 50° to the major one. Rotation from the first girdle to the second one is in a clockwise direction around *c* (looking northwest).

*C1.* The diagram for 200 large quartz grains in a sandy bed in the Peters Creek schist shows a girdle around B. In this, as in some of the other diagrams, the concentrations of *c*-axes fall approximately on two concentric small circles. In C1 most of the major maxima lie on a circle 12° in from the periphery, while a number of the lesser concentrations fall on a circle approximately 45° in from the periphery. The 5 per cent maximum lies at IV-$S_2$. The double 4 per cent submaximum in the northwest quadrant also lies in a IV-$S_2$ position. The small 4 per cent submaximum in the western part of the diagram is at IV-$S_2'$. The center of the large 3 per cent submaximum in the southwest quadrant is 12° from VII-$S_3'$. The 4 per cent submaximum at the bottom lies about 10° from a line joining IV-$S_2'$ and VI-$S_2'$. It also lies in a IV position for $S_2''$, which is a plane 12° counter-clockwise from $S_2'$. The major maximum in the diagram extends inward to include the VII-$S_3''$ position.

*C3.* The small quartz grains show a somewhat disconnected B girdle. The spread of the *c*-axes along the east-west diameter suggests rotation about *a*

(= B′). The 5 per cent maximum lies midway between IV-$S_2$ and VI-$S_2$, but the other concentrations of *c*-axes show better correlation with the later planes, $S_2'$ and $S_2''$.

*C5.* This is a collective quartz diagram of large and small grains. The 4 per cent maximum lies between IV-$S_2$ and VI-$S_2$. The small 3 per cent submaximum at the bottom of the diagram is at IV-$S_2''$, and the 2.5 per cent concentration in the northeast quadrant is at IV-$S_2$.

*D1, D3, D5, and D7* are quartz diagrams of a chloritic quartz-muscovite layer in the Peters Creek schist. These diagrams are for large, small, collective, and elongated grains respectively. They all show a girdle distribution of *c*-axes about B′, which has superseded B as the rotation axis. The girdle appears most distinct in the collective diagram (D5). The positions of the strong 7 per cent maximum and most of the submaxima of the elongated quartz (D7) indicate that many of the grains lie in the $S_2$ plane and are elongated nearly parallel to the B direction. Strong maxima near each end of the trace of $S_2$, which is near the emergence of B, are also present in the collective diagram (D7).

*E1, E3, and E5* are diagrams of large, small, and collective quartz, respectively, in a specimen of Peters Creek schist which has a gneiss-like appearance due to closely spaced syntectonic quartz veinlets. The three diagrams are similar in that all show a disorganized girdle about $a$ = B′, and all their maxima and major submaxima lie in IV-$S_2$ positions. All four IV positions are occupied. The strongest concentration of *c*-axes in the large quartz diagram (E1) is IV-$S_2$ in the northwest quadrant. In both the small quartz diagram and the collective diagram, the strongest concentration is the IV-$S_2$ maximum, southwest quadrant. This appears as a double maximum in the three diagrams. Perhaps the second member of this double maximum occupies a IV-$S_2'$ position which corresponds to a slip plane rotated about 20° counter-clockwise from $S_2$. The diagram suggests a modified monoclinic symmetry, with $S_2'$ (20° from $S_2$) as the symmetry plane.

The mica diagram (E2) indicates a strong point maximum opposite a single slip plane, $S_2$. The distribution of quartz axes so that all four IV positions are strongly occupied tends to confirm that the fabric was governed by slip along $S_2$. After fracturing, the needles of quartz were oriented with their [r:z] edges parallel to the direction of slip; thus the *c*-axes are directed into the four IV positions (Griggs and Bell, 1938). The strong double maxima in the southwest quadrant and the presence of poor girdles indicate that some rotation had taken place, but it is hardly indicated by the micas.

*F1.* This is a partial diagram of 60 large quartz grains (all in the thin section) from a feldspathic quartz-rich bed in the Peters Creek schist. The *c*-axes are roughly arranged in an *ac* girdle. The two strongest maxima lie in the vicinity of the IV and VI positions corresponding to $S_2$ and $S_2'$. The 5 per cent submaximum at the west end of the diagram lies near IV-$S_2'$, and the submaximum just below it is at IV-$S_3$.

*F3.* The maxima and submaxima formed by the small quartz grains lie along an *ac* girdle. They occupy the IV and VI positions related to $S_2$ and $S_2'$, and less frequently $S_3$.

*F5.* The collective quartz diagram is generally similar to F3.

*F7.* This is a diagram of 23 *c*-axes in one large shattered grain. The concentrations can be closely correlated with the maxima positions corresponding to $S_2'$ and $S_3$, rather than $S_2$. This suggests that shattering of the grain post-dates the development of $S_2$.

*G1.* The *c*-axes of 70 large quartz grains (all) in a thin section of feldspathic Peters Creek schist are plotted in this diagram. Only a faint suggestion of an *ac* girdle is present, but there appears to be a *bc* girdle, or perhaps two crossed girdles inclined at a high angle to $S_2$.

*G3.* The smaller quartz grains suggest two crossed girdles more strongly than the large grains. The girdles are inclined to each other about 64°. They are almost symmetrically inclined to $S_2$. The maxima appear to be elongated in the *ab* plane, which may suggest rotation about *c*.

*G5.* The collective diagram of 273 quartz grains also show two crossed girdles, which intersect in the 5 per cent maximum. The nearly horizontal girdle is almost an *ac* girdle, while the vertical girdle rotates around an axis which is inclined about 40° to *c*.

*G7.* This partial diagram of 103 quartz grains, elongated in the *ab* plane, shows two crossed girdles about 85° apart. One girdle is almost perpendicular to *c* and the other is almost perpendicular to *a*. The replacement of B by B′ (*a*) as the chief axis of rotation, was accomplished by a rotation about *c*. This may have formed an *ab* girdle about *c*.

*H1.* The large quartz grains show an *ac* girdle with monoclinic symmetry about an *ab* plane which lies midway between $S_2$ and $S_3$. The maximum and larger submaxima lie in the IV positions corresponding to the several S-planes. However, the 2.5 per cent concentration in the eastern part of the diagram lies at VI-$S_2'$ and VI-$S_2$. The 7 per cent maximum at the top of the diagram lies at IV-$S_2'$ and it is also close to IV-$S_2$ and I-$S_3$.

*H3.* The diagram for the small quartz grains also indicates monoclinic symmetry. The major concentrations of *c*-axes are in the IV and VI positions corresponding to $S_3$, $S_2'$, and $S_3$. The maxima are

double or multiple and show a tendency to be spread out in an *ac* girdle. The girdle, however, is not so good as the one formed by the large quartz grains. The 5 per cent submaxima in the northwest and southeast quadrants lie between IV-$S_2$ and VI-$S_2$, and their position is also common to VI-$S_2$. The large 7 per cent maximum lies inside a small triangle formed by IV-$S_2$, VI-$S_2$, and IV-$S_1$. A small 3 per cent submaximum occurs at the VIII position. The 5 per cent submaximum adjacent to it lies between IV-$S_2'$, VI-$S_2'$, and VII-$S_2$.

*H5*. The collective quartz diagram also suggests a monoclinic symmetry about the *ab* plane. The *ac* girdle is similar to the one in the other diagrams. Major concentrations of *c*-axes fall in the IV and VI positions.

*I1, I3, and I5* are diagrams of large, small, and composite quartz respectively. The specimen was taken from the bottom of a small, tightly folded syncline in quartzitic Peters Creek schist. The mica diagram (I2) indicates that there has been much external rotation about B. The quartz maxima are arranged in somewhat of a girdle which lies 30° or so in from the periphery. The maxima favor positions VI, VII, III, and less frequently, IV. The lineations (B′) pierce the projection sphere about 14° southeast of the upper 7 per cent submaximum in the southeast quadrant of I1. The arc of maxima in the western part of the diagram may thus represent an *ac* girdle about B′. The concentration of submaxima in the southeast quadrant would then be near the VIII position with respect to B′.

*J1* is a projection of 200 *c*-axes of needle quartz in a late syntectonic quartz vein in the slate. The long axis of the needles is oriented almost perpendicular to the walls of the vein. This is interpreted as a growth fabric.

*K1, K3, and K5* are synoptic diagrams of quartz maxima of 3 per cent and higher. The maxima have been rotated so that the *a* axis is at the center of the diagram.

*K3* was constructed from the quartz maxima in diagrams D1, D3, and D7, which are partial diagrams, and D5, which is a collective diagram for rock specimen 150 (Peters Creek schist). Most specimens of Peters Creek schist yield similar diagrams. The maxima are distributed along a cleft girdle about B. The area between the girdles is almost devoid of maxima, and none are present near the emergence of the B-axis. The maxima tend to be somewhat more concentrated near the IV and VI positions, which suggests rhombohedral or pyramidal crystal glide planes parallel to (h01) slip surfaces (Heitanen, 1938). However, the distribution of maxima all along the girdles indicates that intergranular rotation about B was as, or more, important than slip along crystallographic glide planes.

*K5* is a synoptic diagram of quartz maxima in partial diagrams A1 and A3, and collective diagram A5, all of which are from rock specimen 54 (Peach Bottom slate). The quartz maxima in specimen 111 (Cardiff conglomerate) yield a similar diagram. K5 shows the quartz maxima to be concentrated in two crossed girdles inclined to *c* and *b*. Strong girdles suggest that the fabric is due to a crossed strain in which *a* = B′ (Turner, 1948a). The concentration near *a* may indicate a glide direction parallel to *c*.

*K1* was constructed from the 217 maxima in the quartz diagrams A through I. The cleft girdle about B is apparent, but the influence of the slate and conglomerate (A and B diagrams) is seen in the concentration near *a*, which joins the two girdles. Although the quartz maxima are distributed along the two girdles, they tend to be concentrated in the vicinity of the IV and VI positions. One of the concentrations is more peripheral than the others. This may be due to *c* axes lying parallel to (Okl) slip planes (Sander, 1932).

## Summary and Conclusions

The crystalline rocks of this area in the Appalachian Piedmont were deposited as argillaceous and quartzo-feldspathic sediments in a shallow body of water, possibly in Beekmantown or Chazy time. These sediments later formed the Wissahickon and Peters Creek schist. Deposition was followed by slight uplift and erosion, which may represent the Vermont disturbance, the initial uplift of the Taconic Revolution. Slight warping at about this time may have formed a shallow trough-like basin, later to become the Peach Bottom syncline. The coarse Cardiff conglomerate is thought to have been deposited in this basin during Upper Ordovician time. Large quartz pebbles in the conglomerate indicate that the adjacent land was not low-lying. Subsequently, the trough-like basin received quantities of black mud, which were to form the Peach Bottom slate. Sandy intercalations present at the base of the slate may indicate slight oscillation of the shore line.

Pre-Appalachian orogeny may have warped the Peach Bottom syncline and perhaps even folded the rocks, but this cannot be demonstrated. The writer believes the rocks were deformed in a series of overlapping stages during the Appalachian Revolution. The Peach Bottom syncline was compressed by thrusts from the south or southeast, and the beds were thrown into folds parallel to the regional trend of the Appalachians. Tight isoclinal folds were

formed in the slate and the argillaceous beds in the Peters Creek schist. As deformation continued, clastic grains were crushed and stretched. Micas and quartz were recrystallized and reoriented, thus imparting a schistosity or flow cleavage to the rocks. Petrofabric evidence indicates that the micas and quartz grains were subjected to various degrees of rotation about the fold axis, B, and also slippage along S-planes. The quartz fabric continued to undergo change after the micas had achieved their final orientation. In places, quartz shows rotation about B′.

Local inhomogeneity and friction due to drag caused wrinkles or small folds to form along the slaty cleavage planes. Small shear offsets (fracture cleavage) appeared parallel to the axial planes of the folds. Lineations formed at the intersection of fracture cleavage and slaty cleavage. The lineations were parallel to the axes of crenulations and wrinkles.

As slaty cleavage became more pronounced, bedding became less significant. Finally, bedding was almost completely obliterated. It remained where lithologic differences exist. Continued application of stress overturned the slaty cleavage and schistosity planes towards the northwest in the greater part of the area. Thrusts and small low-angle displacements along fracture cleavage planes were directed towards the northwest. These offsets and fracture cleavages intersected schistosity in a lineation parallel to B.

The Cardiff conglomerate as well as competent beds in the Peters Creek schist were greatly compressed and sheared out, but they transmitted a large part of the stresses to the incompetent argillaceous beds. The slate behaved almost as plastic material and responded readily to the deforming stresses. The arching of fold axes occurred in several transverse zones distributed along the length of the syncline. The slate in these zones is badly fractured and contorted, and quartz veins are very abundant. Rotation of fold axes about *c* occurred on a smaller scale in the argillaceous beds in the Peters Creek schist.

The stress direction in major orogenies in Europe, Africa, and elsewhere seems to be meridianal (oral communication from Professor Hans Cloos). If a meridianal stress, resolving itself into two components, one northwest and the other northeast, is postulated for this area, then the compression of the syncline and thrusting towards the northwest could be explained by the northwest component. Wrinkling of slaty cleavage and the formation of cross-folds would then have been formed by the northeast component of stress.

Because the slaty cleavage is overturned to the northwest, and the dip is steep to the southeast, flexuring and cross-folding about *c* imparted a steep east or southeast plunge to the fold axis B′, which lies close to *a*.

Fracture cleavage ($S_4$) intersects slaty cleavage in a lineation which is parallel to B′. The two lineations, B and B′, do not exhibit a uniform relationship as to age. Although B′ frequently transects B, the reverse also occurs. Thus, the two lineations are largely contemporaneous. The slate arched and buckled upward at the same time that it was being compressed, sheared, and displaced along small subhorizontal thrusts to the northwest. Faulting along the north side of the syncline occurred later in the tectonic history.

Contemporaneous subjacent igneous activity is probably closely connected with the orogeny in the area. As the Peach Bottom syncline was not buried at great depth, evidence of igneous activity consists chiefly of abundant syntectonic and post-tectonic quartz veins and a few pegmatites in the Peters Creek schist.

The *ac* cross-joints, which are the most abundant joints, formed during a late stage when the rocks were quite rigid.

Post-deformational, undisturbed Triassic diabase dikes cut the rocks of the region. The history subsequent to the intrusion of the dikes has been one of uplift and erosion.

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# Short Notes

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## PLANATION OF RECENT REEF FLATS ON OKINAWA*

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While engaged in stratigraphic studies on Okinawa in 1947, the writer had occasion to observe rather strong evidence that the present reef flats of that island are the result of comparatively recent wave or solution planation. Evidence indicates that the present reefs were built up several feet higher and that a lowering of sea level resulted in the cutting of the present bench, approximately at high tide level.

In two areas—one along the northern coast of Ie Shima, a small island off the northwest coast of Okinawa, and the other along the southern coast of Okinawa—cliffs of late Tertiary or early Pleistocene limestone form the edge of plateaus that stand 300 to 600 feet above sea level. Along the northern coast of Ie Shima, large blocks from these cliffs have rolled down a steep slope directly onto the reef. Some blocks lie near the reef edge, others probably rolled beyond the reef into deep water. Along the southern coast of Okinawa large blocks from the edge of the plateau can be seen strewn along slopes of the underlying clay formation, and some have come to rest on the present beaches and reef flats. Most of the blocks in this area, after an initial short roll, seem to be creeping slowly down the clay slopes toward the reef.

In striking contrast to the numerous blocks resting on the reef flat, which seem to have reached there very recently, several blocks were seen resting on pedestals of reef rock approximately 5 to 6 feet high. The perched blocks range from the size of an automobile to several times that size. The pedestals on which they sit are composed of algal and detrital reef material identical with and continuous with the present reef surface. The blocks are firmly placed and tightly cemented on some pedestals. On others, such as the one figured (Pl. 1, fig. 1), the blocks are ajar and precariously perched. Still other blocks have fallen and are resting on the reef flat alongside pedestals; the pedestals are apparently in the process of being destroyed and reduced to little more than low humps. There is little to indicate whether the pedestals are undergoing active subaerial or storm erosion at present, or were partially destroyed as a continuation of the processes that formed them during the period of active reef planation. At any rate, some blocks have been lowered from their position on the pedestals to a position alongside blocks that have only recently reached the present reef surface from the edge of the plateau.

No pedestals or blocks were observed in other parts of Okinawa, although reef flats are present along the greater part of its coast. In fact, they are present only where fallen or creeping blocks can still be seen. If the interpretation here made of the pedestals is correct—*i.e.*, that they represent an older and higher reef surface upon which some of the limestone blocks came to rest—then the evidence for higher reefs elsewhere on the island has been destroyed. The pedestals and the supported blocks are a special type of mushroom stack formed by wave abrasion or solution of the original reef after a subsequent fall of sea level. Ordinary mushroom stacks and abandoned nips, although good indicators of former sea levels, do not carry the same implications with regard to former reef levels as the perched blocks of limestone here described do. Mushroom stacks

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can be interpreted as "heads" rising above submerged reefs or lagoon floors and merely nipped later, but there is no implication that the surrounding surface was ever higher than it is now. However, there seems to be no way that a block of limestone weighing many tons could have been transported to the top of a pre-existing "head." The only possible explanation is that it settled on the reef when its general level was about 6 feet higher than at present.

The age of the reef surface represented by the tops of the pedestals is tentatively assigned to the period preceding the last significant change of sea level, a drop of about 6 feet. This period has been correlated (MacNeil, in press) with the so-called "Climatic Optimum" (Flint, 1947, p. 487–499) of glacial geologists botanists, and zoologists and is believed to have been about 4000 to 6000 years ago, when the polar ice caps were smaller and the climate appreciably warmer than now. Evidence for the 6-foot drop in sea level, sometimes called the 6-foot shift, has come from world-wide sources. On the Atlantic coast of Florida and Georgia, bars and lagoons attributable to this stage have been described by the writer (MacNeil, in press). Much of the evidence in the Pacific area consists of abandoned nips and mushroom stacks, many of them far back from the present coast.

Some writers have assumed that the reef flats of the world have been built up to their present level and thus represent constructed platforms. On the basis of the evidence from Okinawa, however, the question immediately arises as to how extensive reef planation has been in comparatively recent times. Particularly the question arises as to how such planation could be detected on low coasts, islands, and atolls where no extraneous blocks of any kind could reach the reef flats. Certainly on Okinawa where the blocks are not present the planation has been complete, and yet there is no evidence of planation in such places. It should be retained as a possibility, therefore, that Recent reef flats anywhere might have been planed from a reef that in comparatively recent time was at least 6 feet higher.

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Plate 1.—BOULDERS ON PEDESTAL REMNANTS

Figure 1.—Boulders of Upper Tertiary Limestone

Resting on pedestal remnant of former higher reef flat, southern coast of Okinawa.

Figure 2.—Boulders on Pedestal Remnants with Plateau Capped by Tertiary Limestone

Large block at left is tightly cemented to pedestal; block to right behind it is ajar. Block shown in Figure 1 in the distance.

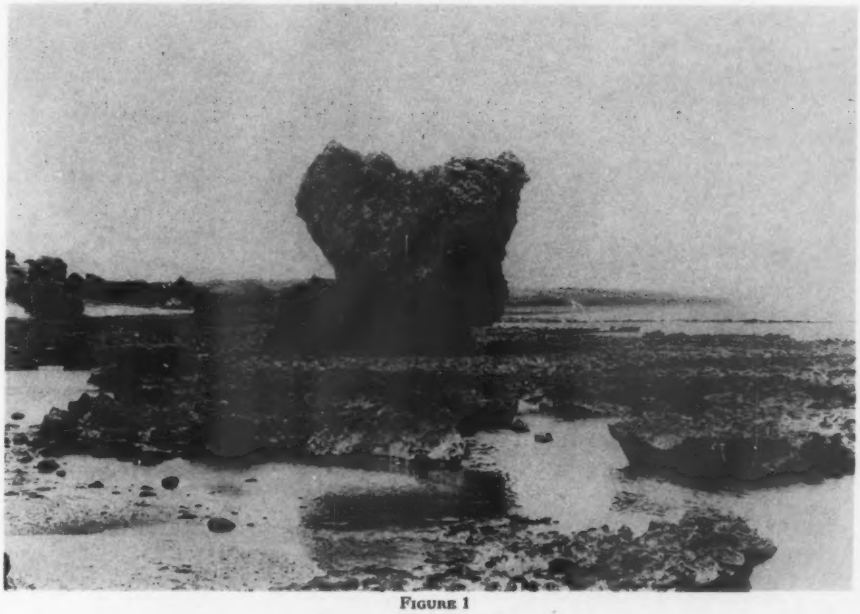

FIGURE 1

FIGURE 2

BOULDERS ON PEDESTAL REMNANTS